
CHAPTER 5 - INSHORE PROCESSES.

Introduction

The Vestfold Hills (Figure 5.1) are an area of ice-free hills, valleys, and fjords about 400 km² in area, on the north-east coast of Prydz Bay. They are bounded by the ice sheet to the east, Sørsdal Glacier to the south, and Prydz Bay to the west. The Sørsdal Glacier follows the contact between the Archaean rocks of the Vestfold Hills and the reworked Archaean gneisses in the northern part of the Rauer Islands. Many other geomorphic features are similarly bounded by major west south west structural features eg. Ellis Fjord, Death Valley (Adamson & Pickard 1986b).

The hills are kept ice free by a ridge of elevated ground which diverts the ice flow to the north and south of the hills (Adamson & Pickard 1986b) and by the high evaporation/sublimation rate of 1220 mm a⁻¹ (Bronge 1989). The area has feeble systems of drainage into closed lakes or tarns, however, there are a few vigorous streams. Precipitation is insufficient to support substantial streams but numerous seepages and minor streams flow through the hills by irregular and opportunistic routes in the summer (Adamson & Pickard 1986b).

The landscape is marked by low rocky hills, till covered valleys, and numerous lakes. Stones and rock surfaces are polished by the action of wind, with small deposits of aeolian sands on the lee side of boulders and rock hills (Hirvas *et al.* 1993).

At the height of the Antarctic summer the temperature of the surface may rise to +20 ° C (Streten 1986), however, the ground remains in permafrost (Hirvas *et al.* 1993, Quilty and Franklin 1997). The permafrost is overlain by an active layer of periodically defrosted sediment from 20 cm to 1 m thick. This layer is typically composed of boulders broken and eroded by frost action and uplifted by frost heaving. Other periglacial features such as frost cracks, patterned grounds, talus fans etc are common (Hirvas *et al.* 1993).

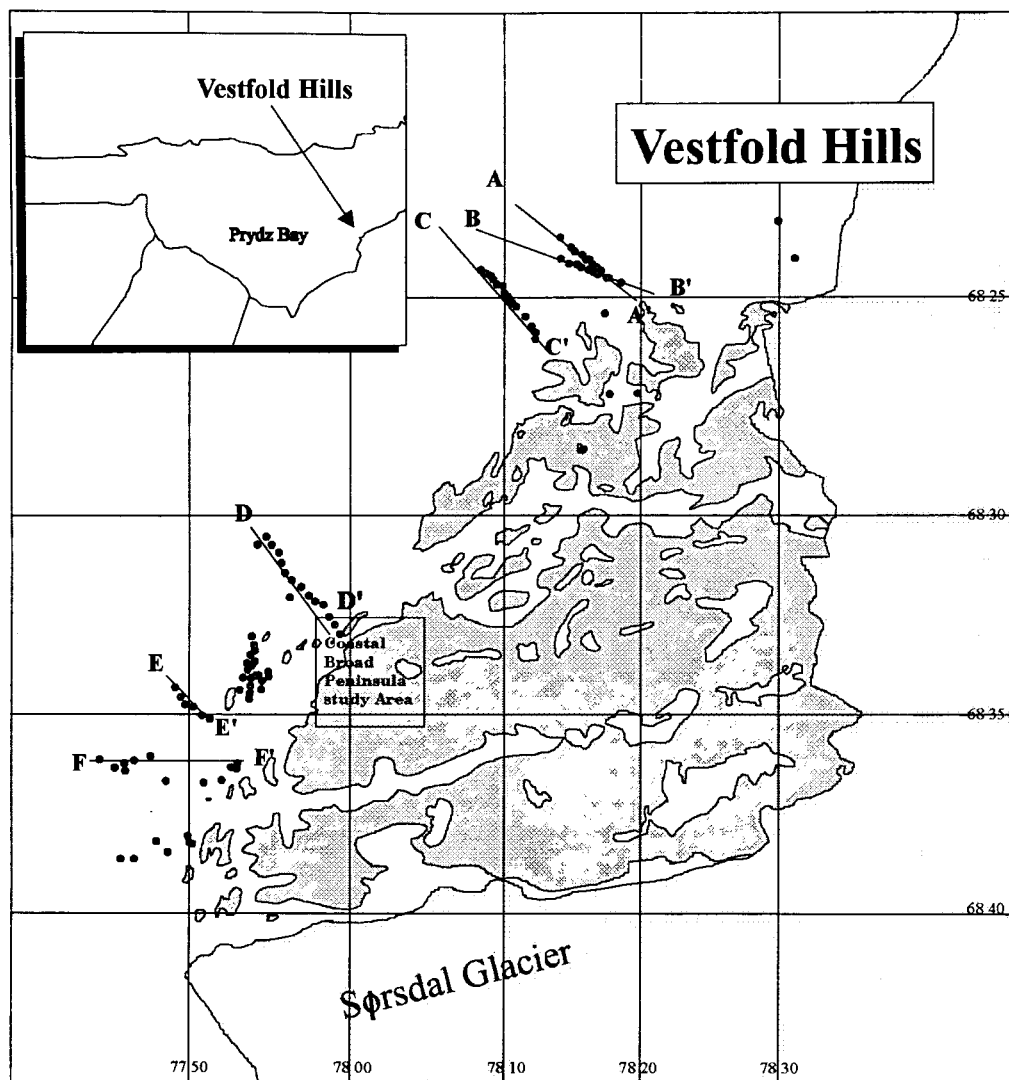


Figure 5.1 - Coastal Broad Peninsula mapping area and location of offshore samples near the Vestfold Hills.

Glacially polished, striated and fractured rocks, roches moutonnées, and ice plucking are common (Adamson & Pickard 1986b). Rock surfaces are etched by weathering, the boulders and rock surfaces are patterned by polyform weathering pits, honeycomb weathering or cavernous weathering (Hirvas *et al.* 1993). This weathering leads to the destruction of the large clasts and subsequent reduction to fine particles (Hirvas *et al.* 1993).

In the Vestfold Hills, roches moutonnée and regional evidence of glacial erosion, “knock and lochan” topography (Hambrey 1994), and glaciated troughs or fjords are evident on a medium to large scale but the only small scale evidence of glacial erosion is seen where glacial ice overflowed basement rock producing striations as described by Adamson and Pickard (1986a)

Samples collected and methods

The Vestfold Hills is represented on the 1:50000 “Vestfold Hills”, 1982 2nd edition sheet. Access to the area is generally by foot. Field mapping was conducted using air-photographs to note outcrop data. One small area was mapped in detail (Figure 5.1, 5.2), namely, Heidemann Valley and Airport Valley from Lake Dingle to the coast. This area will henceforth be referred to as Coastal Broad Peninsula. Sixteen sediment samples and eight drill holes (locations labelled HVDH1-8) were collected at six locations (Figure 5.2). In addition, 125 samples were collected (Figure 5.1) from offshore sites.

Previous Work

The dominant surface sediment covering valleys and hollows is a sediment described by Hirvas *et al.* (1993) as till, occurring as a thin veneer, covering ground moraines, or fields of hummocky moraines. The cover moraine is up to several metres thick and the till veneer is less than 1 m thick. Two discrete till units, separated by shelly sands and clays, were identified by Hirvas *et al.* (1993); the Vestfold or Upper Till and the Heidemann or Lower Till. Marine fossils are present in all of the sediments.

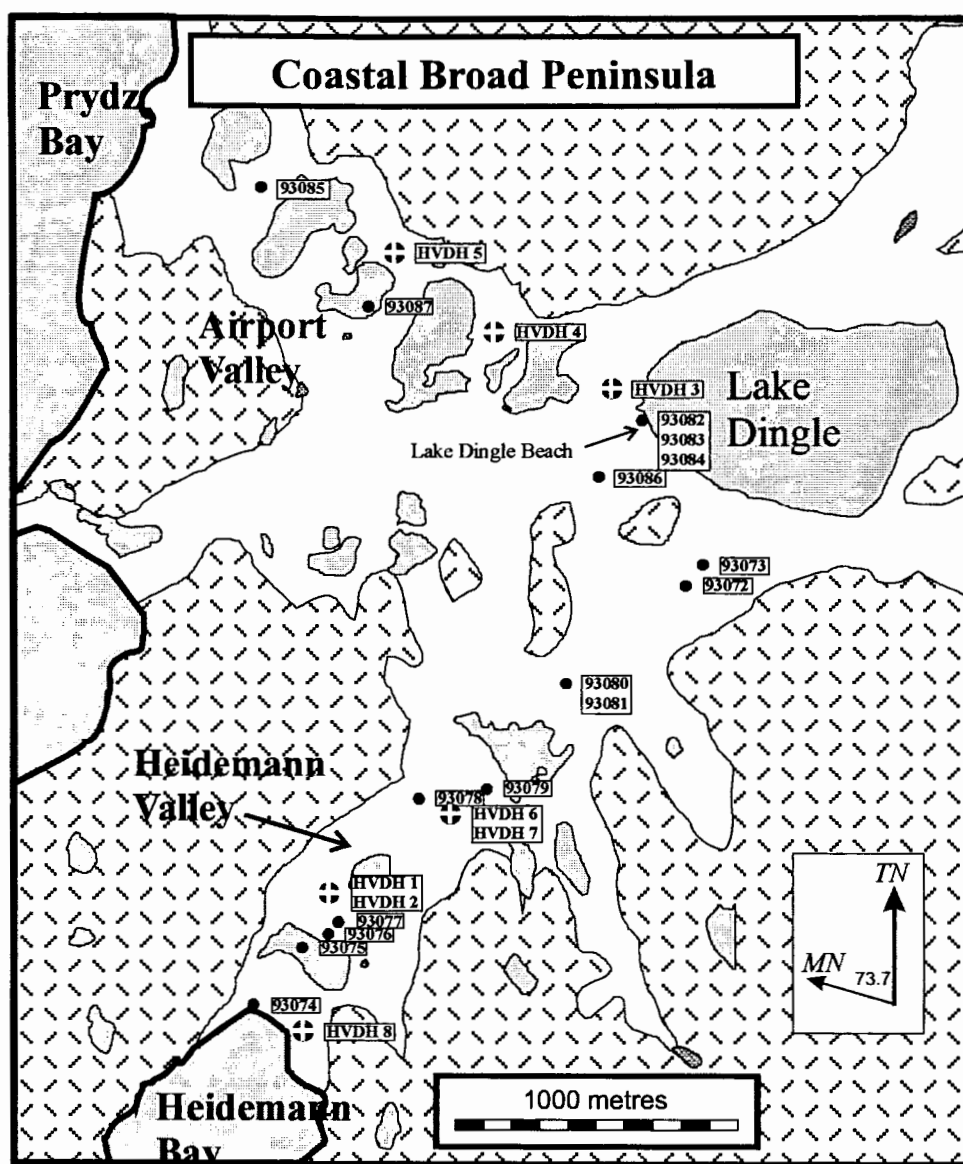


Figure 5.2 - Location of samples collected from coastal Broad Peninsula, Vestfold Hills, Antarctica.

The two "till" units were interpreted as representing two different glacial events (Hirvas *et al.* 1993), however the dates and consequently the timings reported by Hirvas *et al.* (1993) are equivocal. The timing of the deposition of the upper till is between 8 and 300 ka according to Hirvas *et al.* (1993). Further, Hirvas *et al.* (1993) conclude that the upper till probably represents the most recent glaciation when the Vestfold Hills were overrun by the ice sheet and the lower till represents an earlier glaciation (Hirvas *et al.* 1993) although the maximum age presented (300 ka) for the upper till does not support that finding.

Surficial geology and geomorphology

In the Coastal Broad Peninsula area (Figure 5.3) a series of clast supported diamicts forms ridges which cross the valleys normal to the valley axis. The ridges step up in height towards Lake Dingle and are separated by broad flats of matrix supported diamict. In both valleys, but particularly in Airport Valley, the low parts of these flats are overlain by a thin veneer of very fine sand which is deposited by ephemeral lakes which form from the waters produced from large snowdrifts that form on the lee side of the valleys and melt in the summer. These snowdrifts are formed by the contemporaneous deposition of snow and aeolian debris.

RECENT GLACIAL HISTORY

The Cainozoic history of Antarctica is the history of initiation of the largest long-lived ice sheet on earth and this history in the Vestfold Hills is typical of many coastal oases in Antarctica (Adamson and Pickard 1986a). Archaean and Proterozoic basement rocks are thinly covered by Pleistocene and Holocene glacial and marine sediments (Adamson and Pickard 1986a) with 30% of the Vestfold Hills veneered by young material (Adamson and Pickard 1986a, fig 3.3, Gore *et al.* 1996). At Marine Plain extensive early Pliocene marine sediments also occur (Adamson and Pickard 1986a, Quilty 1993).

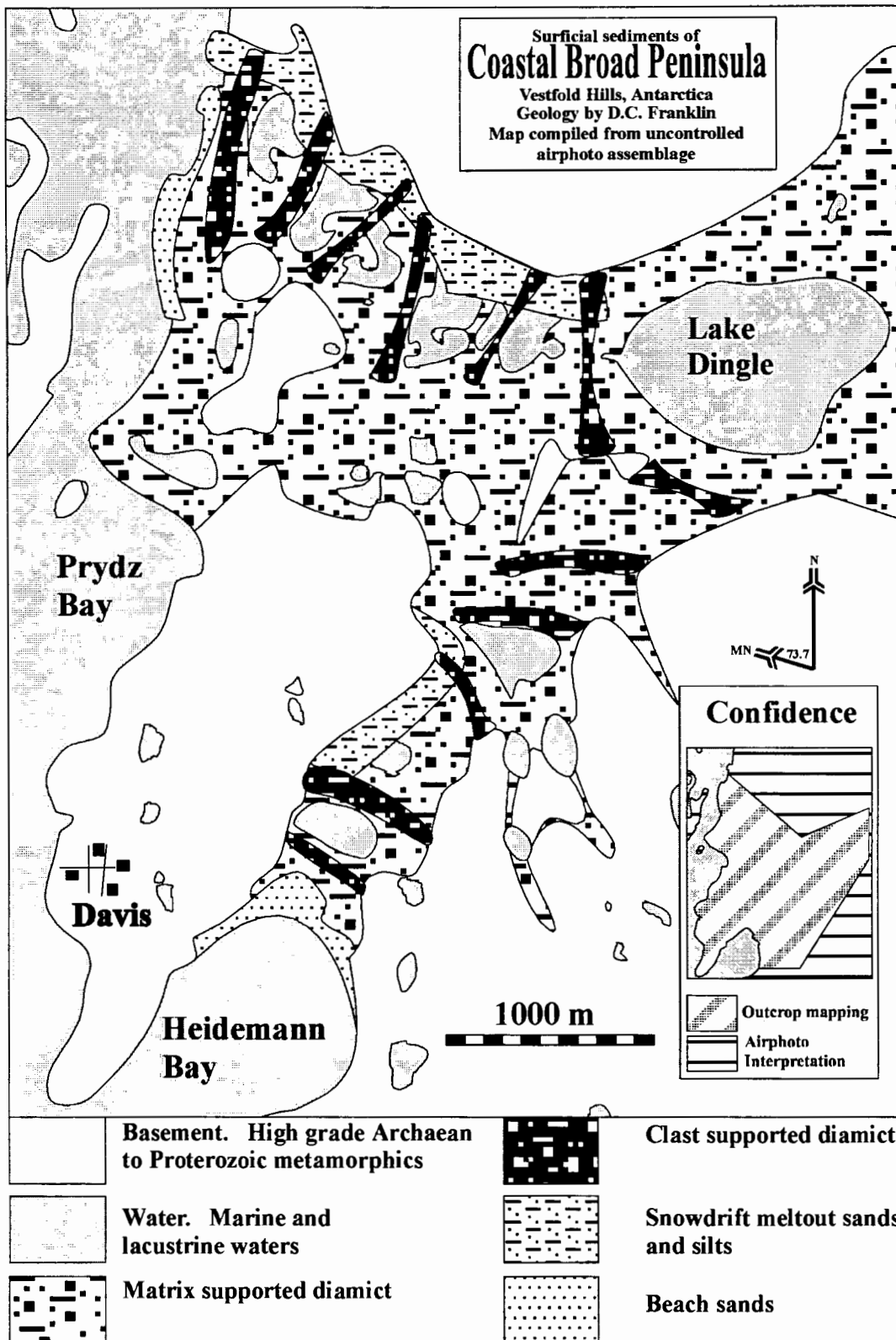


Figure 5.3 - Surficial sediment distribution in Heidemann Valley and Airport Valley, Vestfold Hills.

The Vestfold Hills are presumed to have been covered by at least 1000 m of ice during the Vestfold Glaciation in the late Pleistocene. During the Holocene the ice sheet melted back to its present position and exposed the Hills (Adamson and Pickard 1986a). As the ice receded, additional land was exposed, the sea flooded those low lying areas that were previously covered by ice. During subsequent isostatic rebound the sea retreated and exposed the land. This uplift caused marine inlets, or fjords, to become isolated and form the lakes in the Vestfold Hills. Rates of emergence are currently in the order of 1.8 mm a^{-1} (Adamson and Pickard 1986a, fig 3.20).

Timing for the Holocene Emergence.

The evidence supporting the Holocene emergence of the Vestfold Hills is based on ^{14}C dates of marine organisms in moraines (Adamson and Pickard 1986a, Zhang and Peterson 1985). These data indicate that within the last 8 - 10 ka, the low lying areas of the Vestfold Hills were submerged and presumably subject to many of the same processes that are now occurring in the shallow offshore zone immediately adjacent to the Vestfold Hills. This study, therefore, uses the processes that can be deduced from the patterns of sediments in low lying onshore areas to elucidate those processes. Where possible, processes described in Chapter 4 that have modified the shallow marine sediments after their emergence are mentioned.

SUB-SURFACE GEOLOGY - COASTAL BROAD PENINSULA.

Eight drill holes were drilled over the 1993-1994 summer season near Davis Station to examine the nature and extent of sedimentary facies in support of engineering reports on the feasibility of building an airport runway at one of five possible sites in the vicinity of Davis Station (Quilty and Franklin 1997). As a consequence of the drilling program, however, further scientific study of the sediments in the study area can now be undertaken.

Drilling Results

The eight cores from six sites on coastal Broad Peninsula show very different sedimentary histories for Heidemann Valley and Airport Valley (Figure 5.4). In Airport Valley the sediments are generally high in biogenic silica. The variation in the gravel and sand fraction of these sediments is most likely due to the varying affect of a nearby glacier, supplying debris via ice rafting. As the glacial processes wax and wane in importance, the sediment coarsens and fines in turn. The relative marine influence seems not to change through the cores, except for a peak in HVDH 3 at the 50 cm level. The biogenic silica levels in HVDH 4 are difficult to explain. Placed between HVDH 3 & 5, this core has significantly less biogenic silica, and therefore presumably less marine influence than the other two cores. In other respects, however the sedimentary architecture is consistent for these three sites.

HVDH 1, 2, 6 and 7, from Heidemann Valley are similar, displaying a series of fining upwards units which, based on the occurrence of gravels, sands, and muds have been deposited within a glacial or glaciomarine setting. The major differences however, are that the biogenic silica values are significantly lower than those from Airport Valley and the sediments are generally much coarser in Heidemann Valley. Where Airport Valley silica levels are in the order of 10 to 15% through most of the cores, the values in Heidemann Valley are mostly less than 5% with a few exceptions (core top HVDH 1, 100 – 150 cm in HVDH 2, and 100 cm in HVDH 7).

Any interpretation of the depositional setting of these sediments must account for the marine influence in Airport Valley, the problematic lack of significant biogenic silica in HVDH 4 (Airport Valley), the coarser nature of the sediments in Heidemann Valley, and the apparent lack of marine influence in Heidemann Valley (except for the particular horizons of high silica). The following interpretation of the depositional situation in Heidemann and Airport Valleys is proposed.

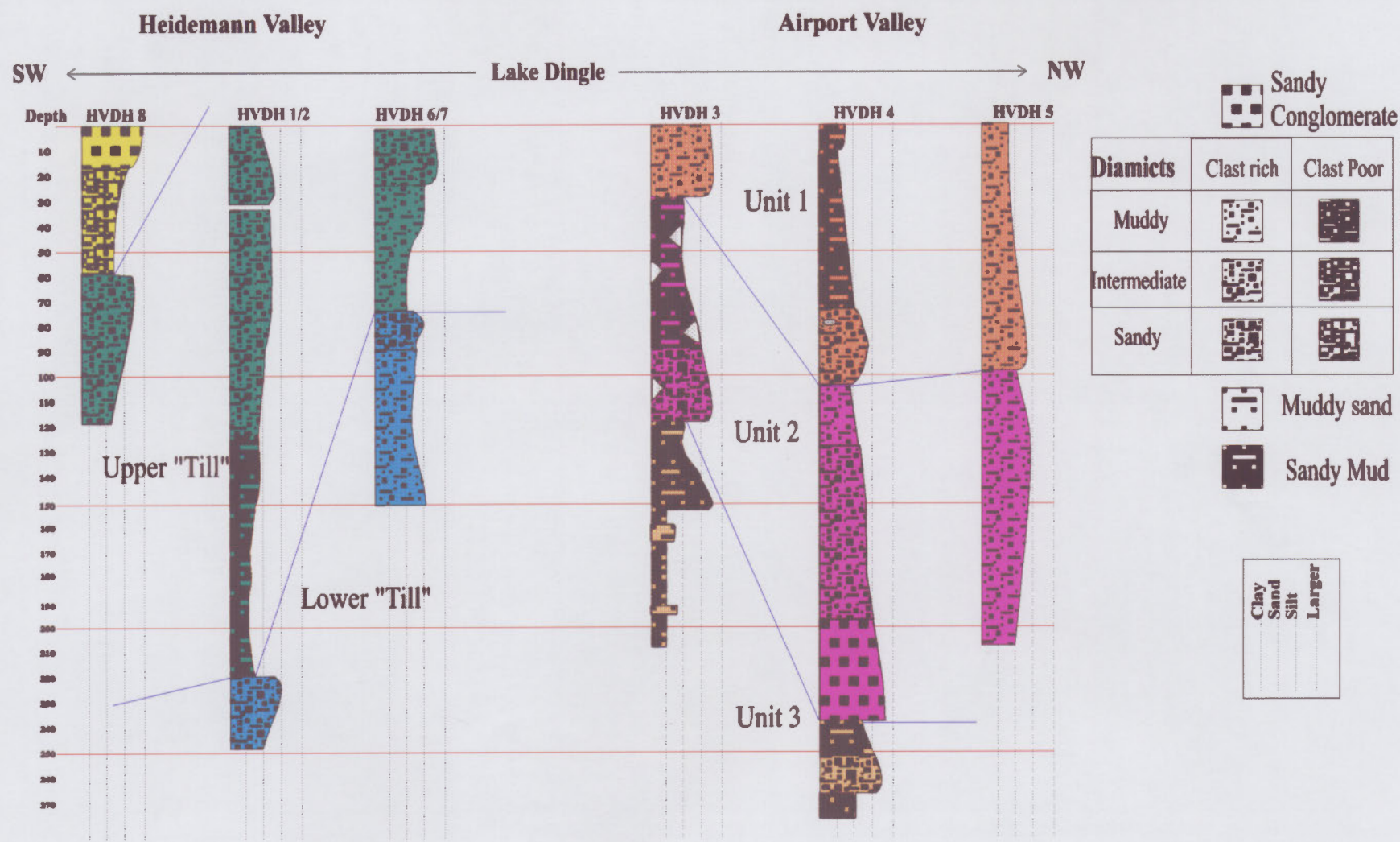


Figure 5.4 - Heidemann and Airport Valley Drill holes. In both cases the seaward end of the section is to the left. Changes in background colour indicate different units.

SEDIMENTOLOGICAL PROCESSES

Deposition in Heidemann and Airport Valleys

Deposition of the boulder fields in Heidemann Valley and Airport Valley occurred in two different regimes and a number of phases. In Heidemann Valley, the deposition of subglacial or waterlain tills interlayered with glaciomarine sediments is difficult to date accurately (Hirvas *et al.* 1993). Dating of a shelly gravel (which was not encountered by the author) between the two subglacial tills has given two ages: one of 50 ka by a ^{14}C date, which is therefore suspect due to the unreliability of the method at ages over about 40 ka; and the second of > 300 ka (Hirvas *et al.* 1993) using a thermoluminescence date which is also suspect (Andrew McMinn personal communication). Thus the dating of the deposition of till unit overlying the shelly gravel is equivocal. Further work is currently underway on this issue with more biostratigraphical and palaeomagnetic data expected to refine the dating of this unit (Patrick Quilty personal communication).

If the uppermost till unit, the Vestfold Till, is older than 50 ka then it must have been deposited at a previous glacial maximum and must therefore have survived the late Pleistocene glaciation intact. This could occur if the last glacial maximum did not result in ice cover in the coastal part of the Vestfold Hills, or if the landforms survived the overrun by glacial ice. The question of the survival of small scale landforms under glacial ice was investigated by Kleman (1992), who hypothesised that if landforms that predate ice-sheet burial can survive subglacial conditions and the possibility is not recognised, then erroneous conclusions about ice-sheet dynamics and landscape development may be drawn. Kleman (1992) concludes that, as basal sliding cannot occur under cold-based conditions (Boulton 1972, Hughes 1973, Drewry 1986) and because of the evidence for preservation from a variety of regions, settings, environments, and scales, that such preservation is possible.

Alternatively, the possibility that glacial ice did not extend as far as the current coastline was raised by Adamson and Pickard (1986a), suggesting that the boulder fields and depositional forms would have remained completely undisturbed during the last glacial maximum. Were this the case, isostatic depression would have seen this area submerged and undergoing glaciomarine sedimentation. Indirect evidence that this was the case is seen in the much more recently glaciated surface in Heidemann Valley to the east of Lake Dingle (Quilty and Franklin 1997).

If there was an active glacier in Death Valley during the last glacial maximum, it is possible that the ridge surrounding Lake Dingle on its seaward side is a grounding line moraine. Although measurements were not taken there are a number of features of this ridge that suggest it may have been a grounding line moraine. Firstly, the ridge joins high points of basement rock which could have formed a stable sill for the glacier similar to the tidewater glaciers described by Powell (1981). Secondly, the back of the ridge forms a single reasonably shallow slope dipping upstream. Finally, an esker like feature was noted on the upstream side of the south-western extension of the ridge. This flat topped feature was oriented normal to the ridge, was approximately 4 m high and composed of well sorted, very coarse gravel.

The results of this study can assist in choosing between these two possibilities: subglacial deposition of Heidemann Valley "tills" (Hirvas *et al.* 1993) as opposed to glaciomarine deposition (Adamson and Pickard 1986b). The very low, yet measurable amounts of biogenic silica, visible forams, radiolaria, molluscs, and glass sponges (Hirvas *et al.* 1993) in all of the sediments in Heidemann Valley suggest some marine influence in their deposition. In addition, a spike of high silica content at about the 100 cm level at the base of Hirvas *et al.*'s (1993) Vestfold Till unit probably indicates a significant marine influence. Such would not be expected if the units were sub-glacial. Thus, Adamson and Pickard's (1986b) theory that the glaciation did not extend to the current coast but only as far as Lake Dingle is supported by this evidence. The overwhelming glacial influence may be explained by waterlain deposition

beneath a floating ice tongue in Heidemann Valley and the periodic collapse of that ice tongue could explain intervals of high marine influence.

Regardless of the depositional situation in Heidemann Valley, the sediments in Airport Valley are very different being high in biogenic silica. The sediments are clearly influenced by both marine biogenic and glaciomarine sedimentary processes.

This evidence is interpreted as indicating a depositional environment with a minor glacier occupying Death Valley, decoupling from the sea floor around the seaward side of Lake Dingle and terminating in a floating ice tongue or minor ice shelf occupying Heidemann Valley during deposition of the Heidemann Valley sequence (Figure 5.5). There was little or no extension of a permanent floating ice mass over Airport Valley. Whether the two sequences; Heidemann Valley and Airport Valley, were deposited at the same time is not known and accurate dating is required.

Formation of boulder armour

The surface of all of the sediments in Coastal Broad Peninsula is armoured by a layer of boulders which is mostly one clast thick but up to two metres thick in the vicinity of the beach ridges. This armour is formed by freeze thaw processes in the active layer above the permafrost. The abundance of fines in this layer assists the frost action (Hirvas *et al.* 1993) in the upward migration of large clasts.

Beach ridge construction

Ridges occur along both Heidemann Valley and Airport Valley. These ridges are arcuate features normal to the central long axis of the valleys, concave on the seaward side. There are five of each of these ridges between the ridge surrounding Lake Dingle and the sea in each valley. Ridges are approximately 2 to 3 m in height above the surrounding landscape and they are tens of metres wide, and formed of a clast supported, boulder rich diamict.

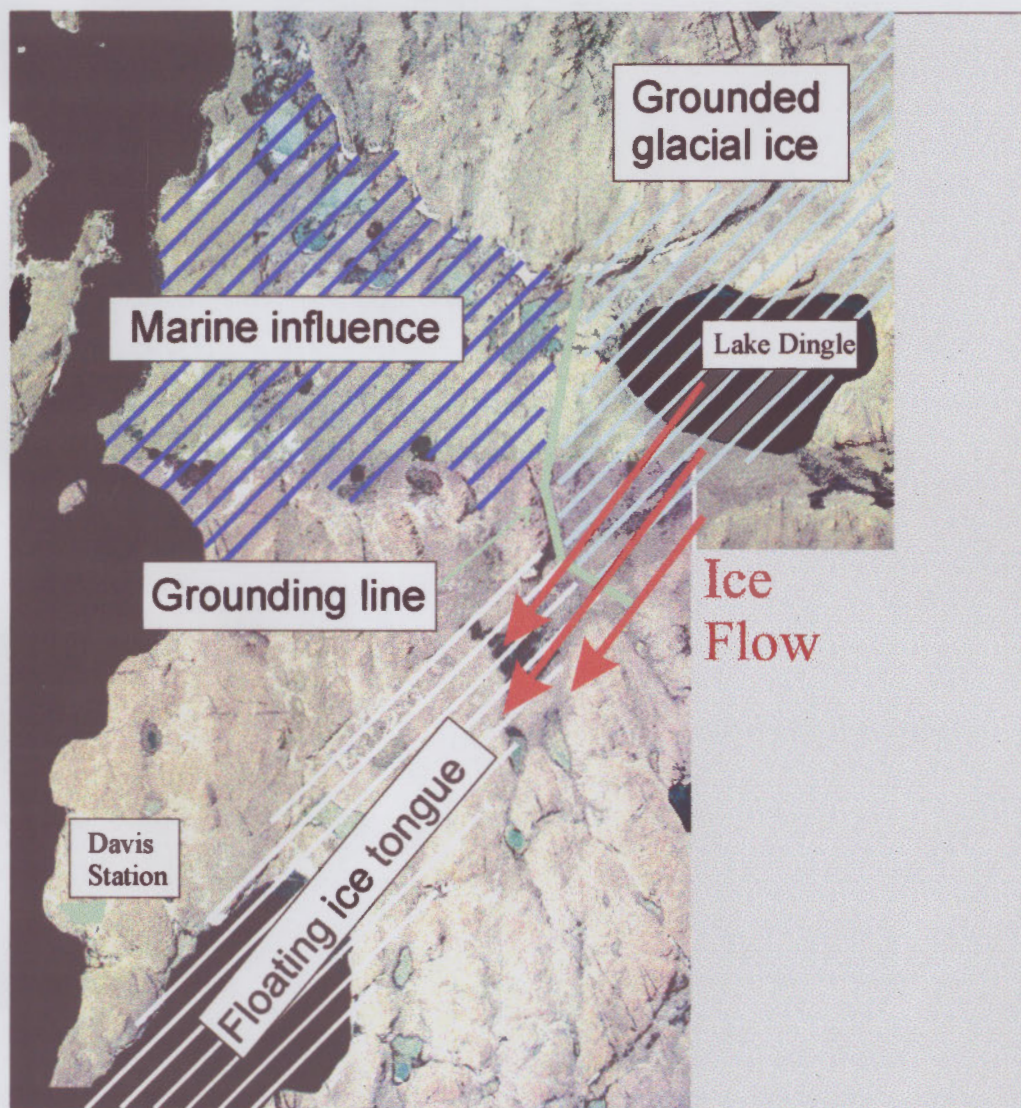


Figure 5.5 - Hypothetical representation of ice dynamics during the time of deposition of the Heidemann Valley units.

The up-valley curves on these ridges and the cross-sectional morphology suggest these are not grounding line or terminal moraines and it appears that they are formed by the action of sea ice (Figure 5.6). As the land surface emerges from the sea, new sections are exposed in the intertidal area. Sea ice freeze-in incorporates smaller clasts and fines which are removed during the summer and deposited elsewhere when the ice melts. During the winters, the sea ice is moved up and down by the tides and tamps down the sediments with which it is in contact. During this study, the sea ice was consistently measured at 2.1 m thick in the vicinity of the Vestfold Hills and so the depth to which this process occurs is 2.1 m plus the tidal range (60 cm).

The cross-sectional morphology of these landforms (Figure 5.7) is best seen at the most recent ridge in Heidemann Valley (Ridge A). From the seaward side the surface slopes gently upwards and abruptly flattens out to a horizontal or slightly reverse slope. This flat zone terminates against a ridge which is approximately 2-2.5 m high and up to 50 m wide. The back of this feature then slopes down into a depression which dips below the level of the flat bench. The surface then slopes gently upwards again to the next bench and ridge formation (Ridge B).

The probable process that forms these ridges is the lateral expansion of the sea ice during its formation in early winter, pushes against the sediments and forms the ridges which cut across the valley. According to Adamson and Pickard (1986a), the Holocene deglaciation of the Vestfold Hills occurred in steps with slow melt back of the ice sheet punctuated by accelerated melt and associated increase in uplift rates. Such a situation may explain the stepped beach ridges in Heidemann Valley and Airport Valley.

Similar laterally extensive landforms, identified as intertidal boulder pavements, are formed in Ice Bay in the Gulf of Alaska by the tamping action of glacial ice and winnowing by marine processes (Eyles 1994). They appear much more mature than the benches in Heidemann Valley; however they vary from poorly developed to very well developed (Eyles 1994). Pavement formation occurs as

marine lag surfaces are abraded by the grounding of icebergs and seasonal pack ice as seen in the South Shetland Islands (Araya and Herve 1972) and Vestfirðir, Iceland (Hanson 1983, 1986). Although the mechanisms here are different, the concept that ice and marine processes in concert may cause the boulder benches in Heidemann Valley is supported. Critical differences are that the Heidemann Valley structures are built by sea-ice, not glacial, action.

Sandy and gravelly beaches are found at present sea level in many parts of the Vestfold Hills. Many are rich in garnets from the local metasediments. There are very few sandy beaches above the waterline however. It is likely that beaches have been forming throughout the Holocene emergence of the Vestfold Hills and so the lack of preserved beaches needs to be addressed. Old beaches do occur along the shore of a small unnamed bay (GR 840825 Vestfold Hills 1:50000 2nd Edition) which is almost cut off from Ellis Fjord by a large lateral moraine which was supposedly emplaced by the lateral expansion of the Sørsdal Glacier during the Chelnok Glaciation. Samples collected from this area from both the boulder ridges and the intervening sandy areas are similar; however, the sandy unit is better sorted and has a coarser grain size. This is interpreted as sorting due to wave action while the fine skewed sands from the boulder ridges are closer to subglacial till sand sorting and may represent less wave action. The preservation of these ridges may be due to the presence of the large lateral moraine which prevents removal of sands by sea-ice freeze in or wave action.

Observations of the currently active beach environments show that common beach forming processes are occurring along the coastline of the Vestfold Hills. The sands originate from the weathering of onshore metasediments and volcanics as discussed in Chapter 4, and are blown into the inshore environment by the persistent offshore winds. Wave action during the summer pushes these sands and gravels onto the shore face. Tidal currents are gentle when there is no sea ice and it is unlikely that much sand is removed by them.

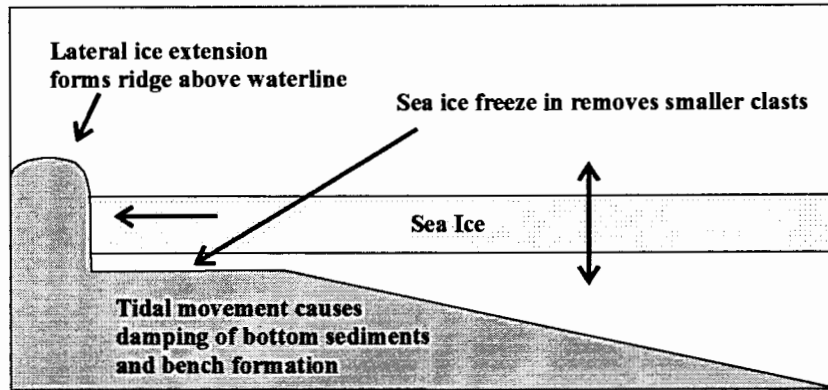


Figure 5.6 - Diagrammatic representation of the formation of beach ridges and benches by sea-ice.

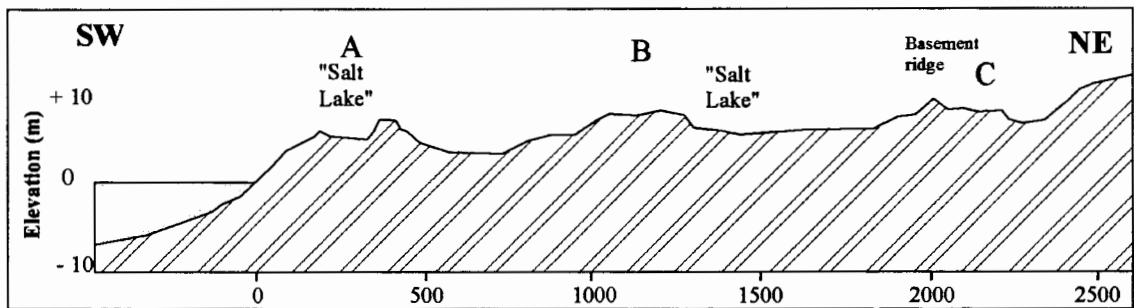


Figure 5.7 - Profile along Heidemann Valley. Heidemann Bay is to the left. Ridge A is the most recently formed ridge and a large salt lake has formed on the bench. Ridge B is an earlier example while Ridge C is composed of basement.

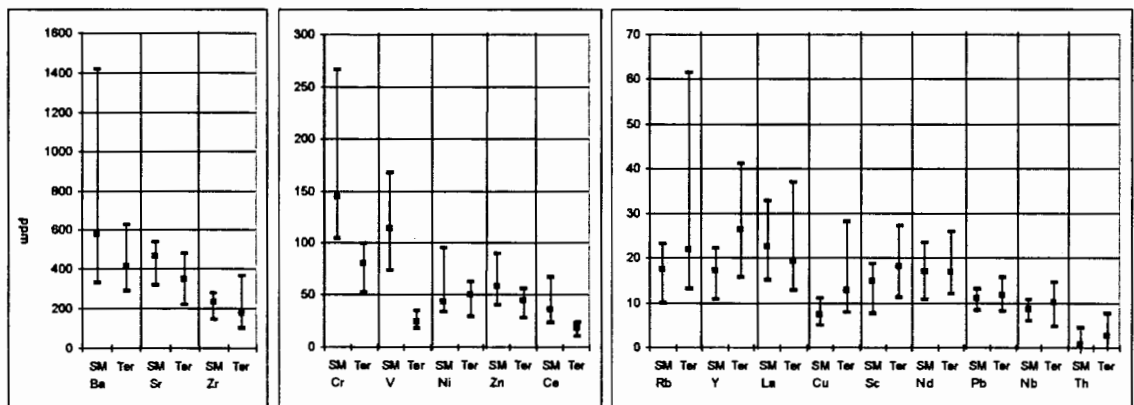


Figure 5.8 - Trace element geochemical results. For each element the plot shows shallow marine samples to the left and terrestrial samples to the right.

It is likely that beaches are not preserved because once they have emerged from the sea and dried, they are remobilised by the wind. Sea ice freeze in probably also accounts for some of the removal of material.

Trace Element Geochemistry

Indirect evidence of the close relationship between onshore and offshore sediments is seen in the trace element geochemical signatures of the two sediment stores (Figure 5.8). Sc, V, Cr, Ni, Y, and Cu are depleted in sediments offshore from the Vestfold Hills in comparison to onshore sediments but the ranges of these elements overlap substantially. Nb, La, Ce, Nd, Pb, Zr, and Th show no significant difference between onshore and offshore sediments. Ba, and Sr are slightly enriched in offshore sediments due, probably, to the concurrent deposition of BaSO_4 as a biproduct of bacterial productivity, and strontium as a replacement element in biogenic calcite.

Aeolian deposition of sediment in the inshore environment

The removal of fine grained material from exposed terrestrial sediments and the subsequent transport into the shallow marine environment results in terrestrial sediments which are coarse grained with respect to the primitive sediments (Figure 5.9), and shallow marine sediments which are relatively fine grained. Gravel proportions do not differ significantly between the groups (Figure 5.10). The enrichment of fines in shallow marine samples is due in part to the aeolian input of fine grained material from exposed terrestrial sediments and in part by the addition of the remains of planktonic organisms (Figure 5.11).

Further evidence of the aeolian transport of sediments across the coast and into the shallow marine sediments is seen in the skewness of shallow marine sediments with respect to terrestrial sediments (Figure 5.12). Sediments become strongly coarse skewed with decreasing mean grain size with marine sands more coarsely skewed than terrestrial sediments. This apparently problematic situation is due to the aeolian transport of fines across an unchanging substrate of coarse sediments.

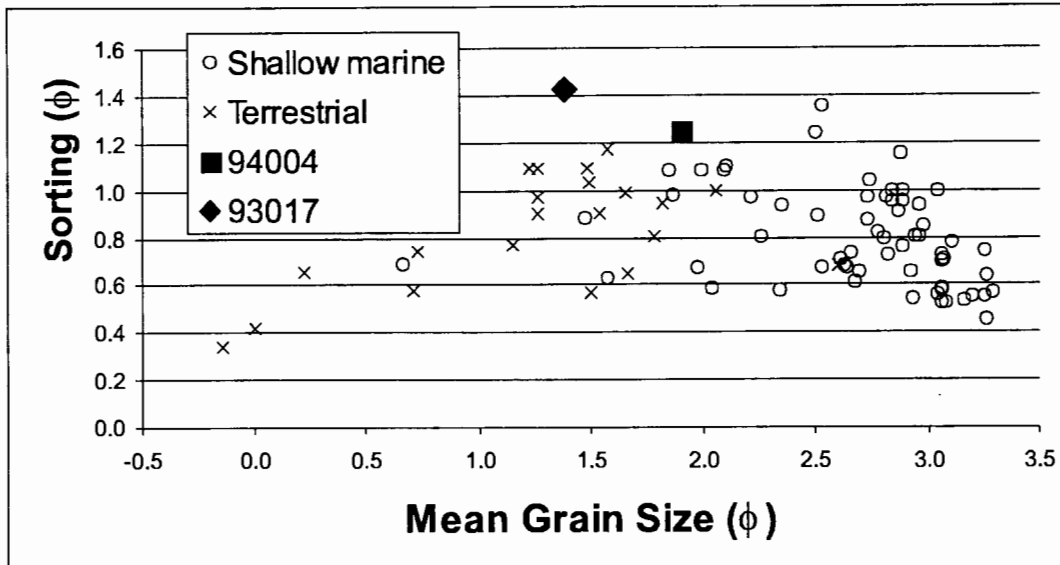


Figure 5.9 - Bivariate plot of mean grain size (ϕ) against Sorting (ϕ). Shallow marine samples group to the fine end while terrestrial samples tend to be coarser. The groups overlap significantly however.

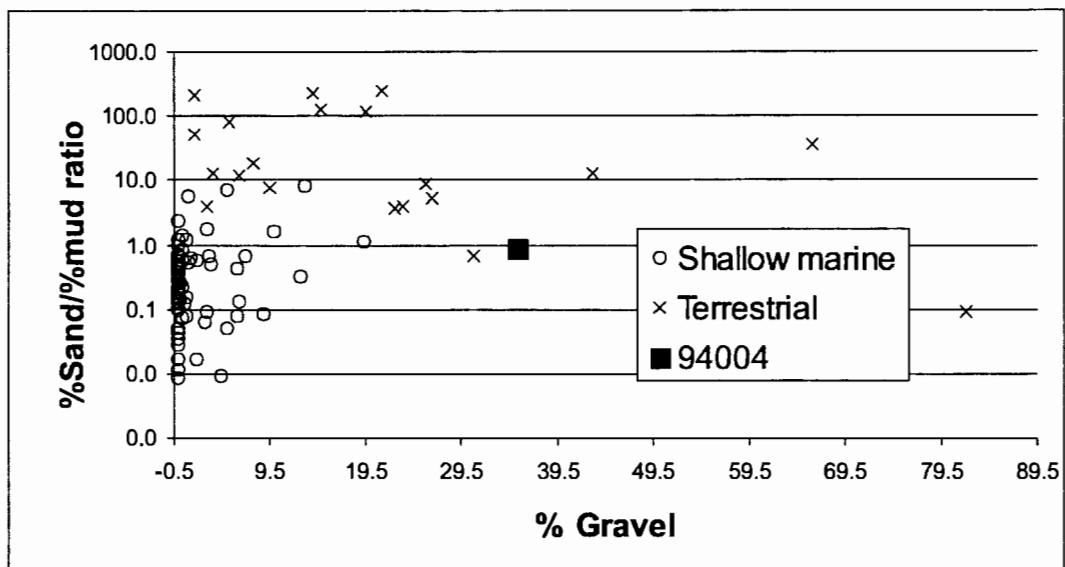


Figure 5.10 - Bivariate plot of % gravel against the sand/mud ratio. Terrestrial samples generally contain more sand but the groups overlap considerably.

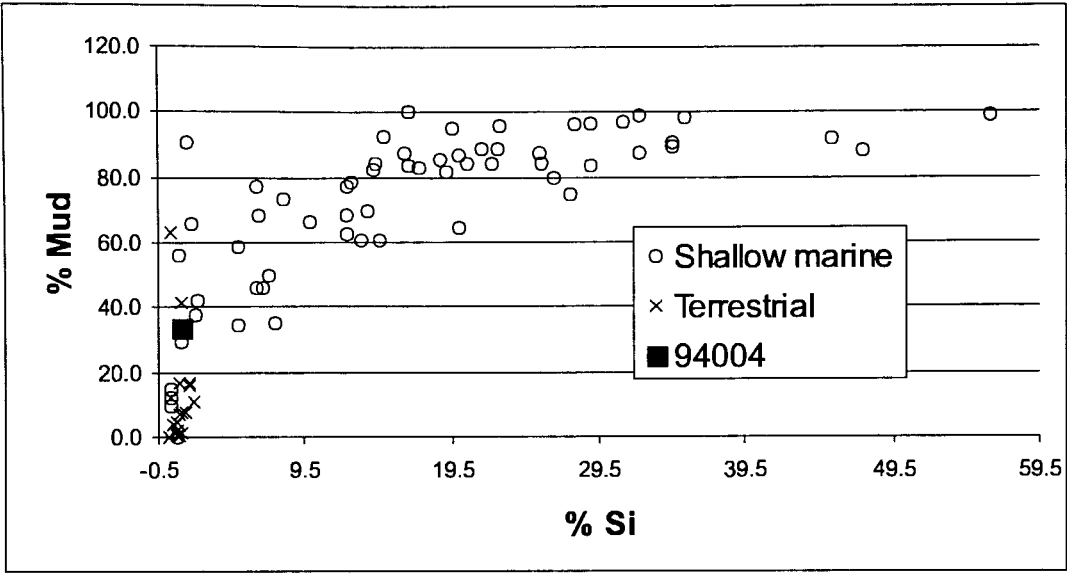


Figure 5.11 - Bivariate plot of % Si against % mud. Shallow marine samples are enriched in Si compared to terrestrial samples.

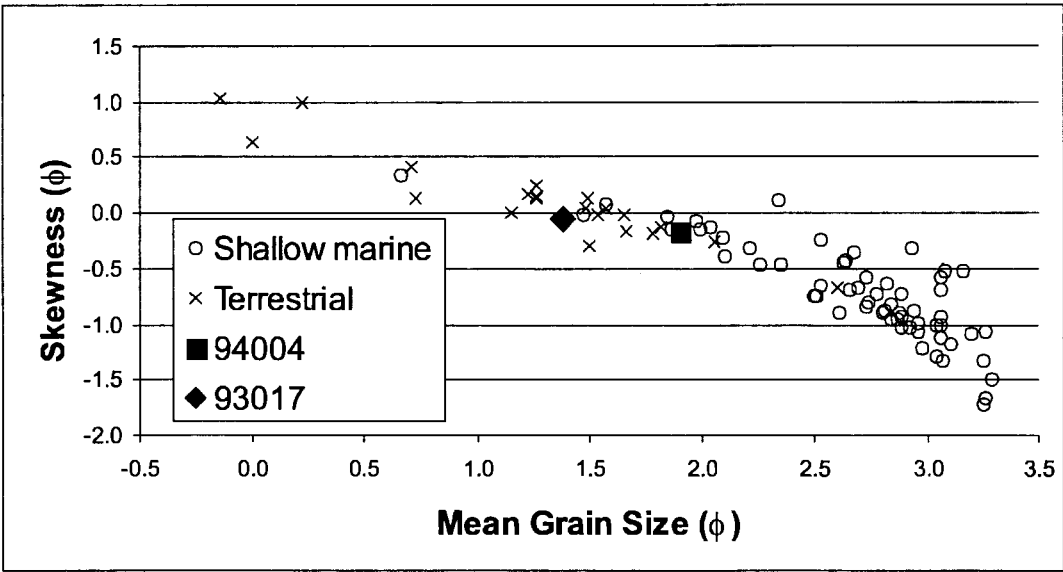


Figure 5.12 - Bivariate plot of mean grain size against skewness. Shallow marine samples are fine grained and coarser skewed.

As the distance over which aeolian transport occurs increases, the mean grain size of the transported fraction decreases while the substrate grain size characteristics remain unchanged. The relative effect of the *in situ* sediments on the grain size distribution of the whole sediment is therefore increasingly strong coarse skewing as the average grain size of the transported material, and consequently the whole sediment, decreases.

In summary, aeolian transport preferentially removes fines from terrestrial sediments (Figure 5.13). This material is transported to the shallow marine environment, reworked by wave action, depleted (non-preferentially) by sea ice freeze in, and augmented by fine biogenic particles. This results in a finer grained, better sorted sediment than the original terrestrial deposits. Relative sea level drop can then re-expose the shallow marine sediments, making them again available for this set of processes

Biogenic input

Biogenic carbonate is uncommon in the offshore and onshore surface sediments sampled and was not found in the onshore sub-surface sediments sampled, despite being reported by Hirvas *et al.* (1993). The maximum value encountered was a sediment with 22.82% carbonate and only 14 samples contained sufficient carbonate to measure. The significant presence of biogenic silica in shallow marine sediments is expected, with up to 56% of one sample composed of silica. Many of the high values of biogenic silica in terrestrial samples are from downcore samples in the three drill holes from Airport Valley on Coastal Broad Peninsula suggesting a marine influence in these sediments.

As with most marine areas in Prydz Bay, samples collected contained abundant biota from large bivalve molluscs and annelid worms to echinoids, holothuria, and foraminifera. This biota is responsible for significant bioturbation, and in the shallow marine environment is probably responsible for the mixing of material being deposited now with older material beneath the surface of the sediment water interface, although the ability of the biota to move large clasts is very limited.

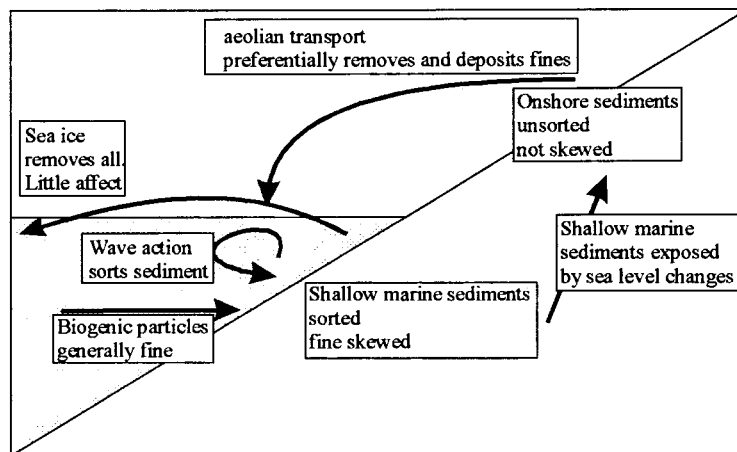


Figure 5.13 - Aeolian transport of unsorted terrestrial material into the shallow marine environment, and subsequent reworking processes.

OFFSHORE SEDIMENTS

Six transects were sampled in the nearshore areas of the Vestfold Hills. These transects were oriented approximately normal to the coast. The sediment distribution is complex with wide variation on a small scale, but in general, SMO occupies the topographic depressions, while basement and till form the highs (Figure 5.14). As with onshore topography, the relief is high in the north becoming more gentle in the south although without adequate bathymetry this is difficult to quantify.

The lack of a detailed bathymetric chart of the area sampled makes interpretation of the sediments difficult, however, it is possible to extrapolate the general topographic pattern to areas offshore. A generalised map (Figure 5.15) of the sediments encountered shows SMO infilling the topographic depressions and being interspersed with high outcrops of crystalline basement or till. More sampling would enable the patterns of surficial sediments to be better elucidated. This pattern suggests that the sediment is only in low lying areas and broadly agrees with the distribution pattern observed in onshore sediments with sediment free ridges and valleys infilled with glacial or glaciomarine sediments.

CHAPTER SUMMARY

Shallow marine depositional processes near the Vestfold Hills are dominated by the deposition of clastic material by aeolian transport and, less importantly, the *in situ* deposition of biogenic carbonate and biogenic silica.

Sediments sampled from low lying terrestrial areas are remarkably similar to those in the shallow marine area. Biogenic carbonate and silica concentrations are higher in shallow marine samples, while grain size statistics of the sand fraction of samples differ significantly and represent the evolution of grain size characteristics as sediments are entrained by the wind onshore, and transported offshore.

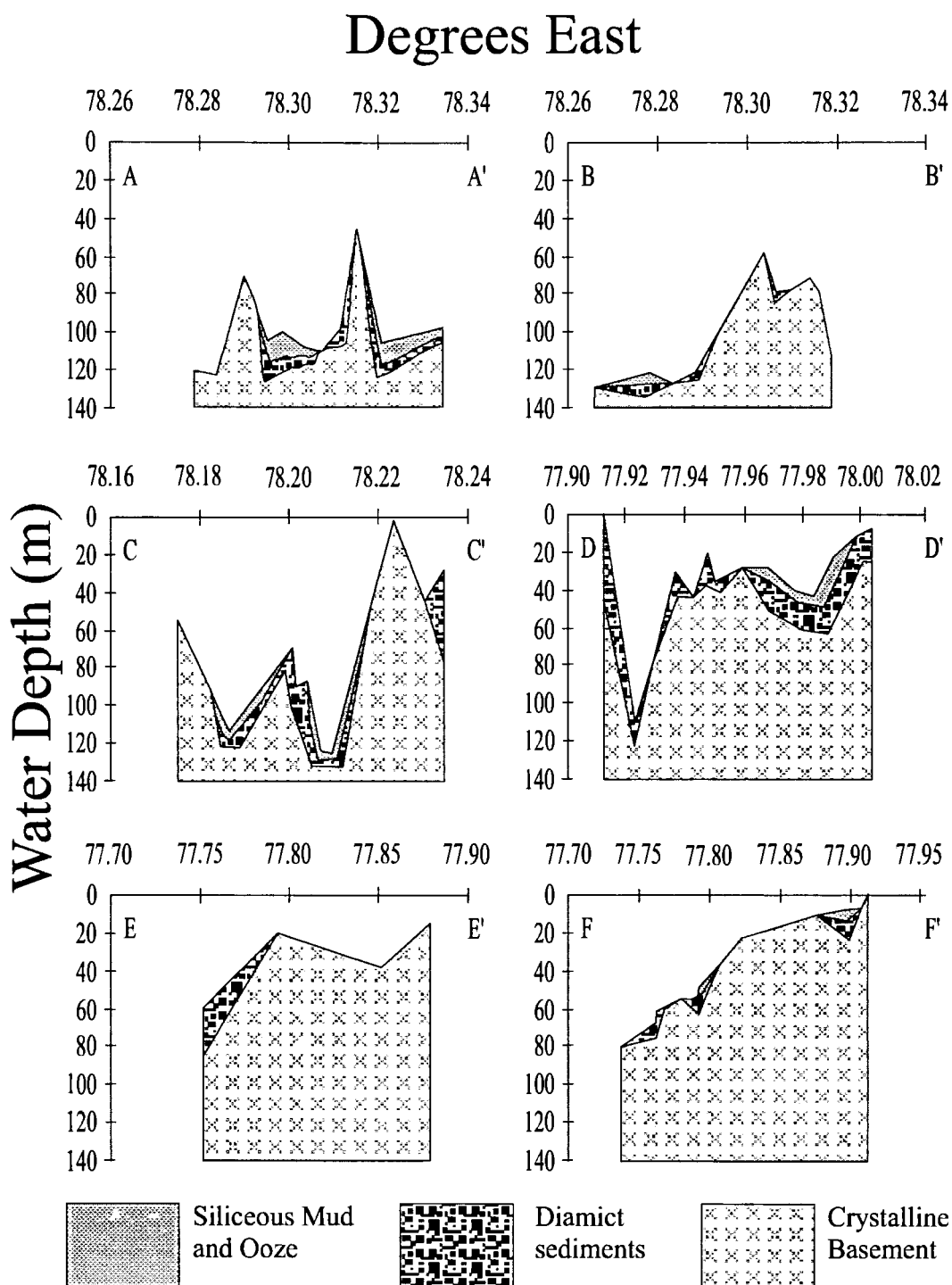


Figure 5.14 - Bottom sediments and proposed profiles from 6 offshore transects near the Vestfold Hills, Antarctica.

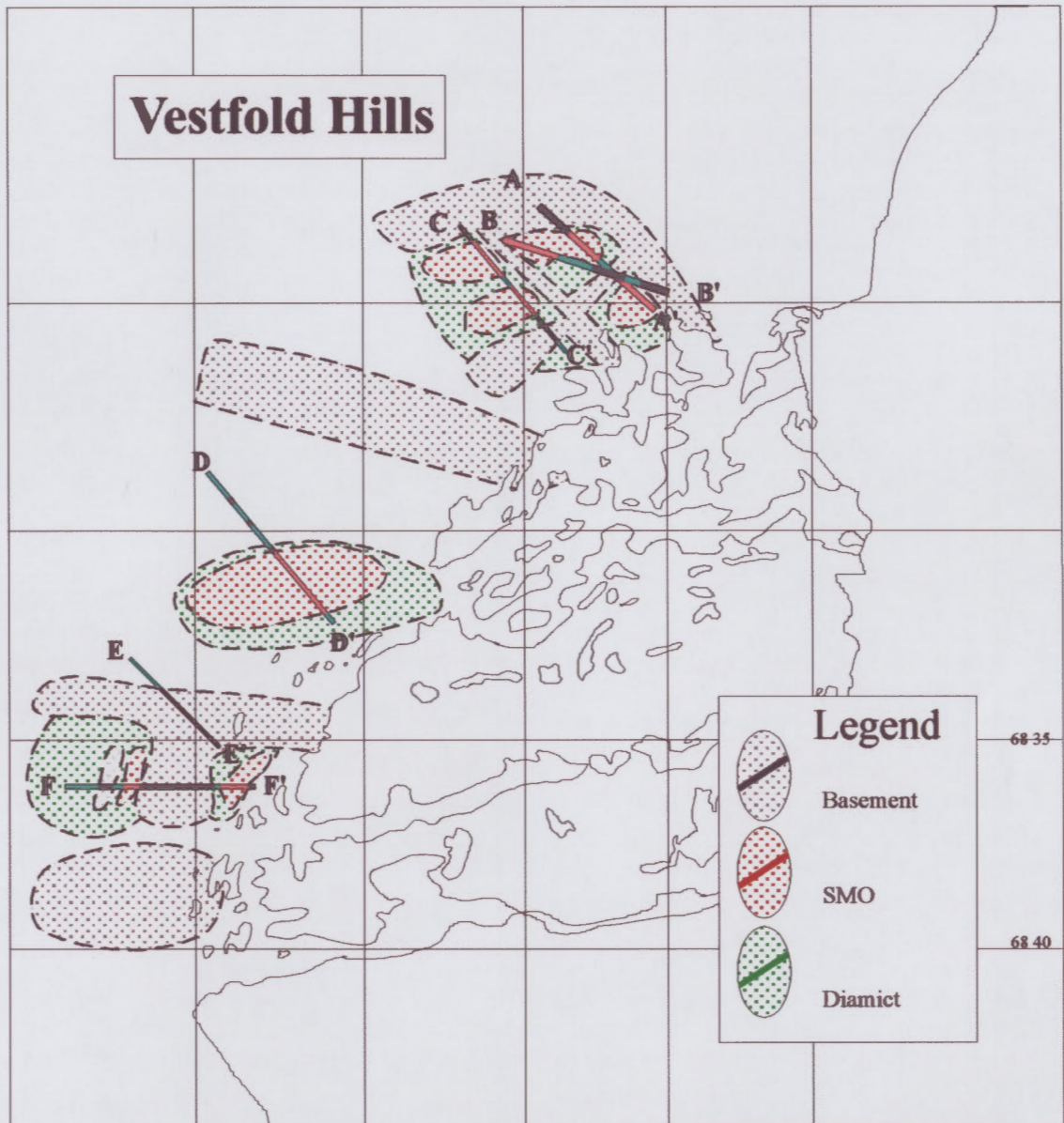


Figure 5.15 - Proposed sediment distribution deduced from onshore sediment architecture and offshore transects.

Because the offshore areas sampled are contiguous with onshore areas, and there is ample evidence of the marine influence in the onshore areas, the onshore distribution of sediments is representative of the sediment distribution offshore. Onshore areas provide a model for shallow water depositional environments near the Vestfold Hills (Figure 5.13).

Sediments are blown from the Vestfold Hills to sea-ice or open sea where they are augmented by the *in situ* production of biogenic carbonate and silica and strong and persistent bioturbation mixes this with older underlying sediments.

Sea ice action is important in the production of benches near beaches. Long periods of static sea level produce beaches which are subsequently blown away when above sea level or bioturbated when below sea level while the variable rate of isostatic uplift due to deglaciation during the Holocene has formed a series of these benches.

Contrary to the results of Hirvas *et al.* (1993), Heidemann Valley sub-surface sediments are waterlain tills interlayered with glaciomarine diamicts which were deposited beneath a floating ice mass which was subject to periodic collapse. Airport Valley sub-surface sediments are wholly glaciomarine. Shelly intervals in these sediments represent surface accumulations of biogenic carbonate deposited during periods when the ice tongue collapsed.

CHAPTER 6 – OFFSHORE PROCESSES

Introduction

There have been a number of studies on the sedimentology of Prydz Bay. Quilty (1985) identified three major sediment types; those dominated by silt and clay size fractions, terrigenous sands with little biogenic component, and coarse sediments with biogenic carbonate. Within his study area, Quilty (1985) observed that the coarse sediment which was high in carbonate was mostly located in the northern part of Prydz Bay and the Mac. Robertson Shelf, mostly on, and to the north of, the shelf break. Within Prydz Bay proper, he observed that the main sediment was the “dominantly silt and clay” sediment, with or without a significant biosiliceous component. This pattern was again observed by Franklin (1991, 1993), with areas of high carbonate (up to 50%) on, and to the north of the shelf break, and “sandy diamict” and “massive mud” sediments within the bay.

The Ocean Drilling Program drilled 5 holes in Prydz Bay (Barron *et al.* 1991), however much of the Quaternary sediments obtained were disturbed (O’Brien *et al.* 1993). The Holocene section in ODP Hole 740A (Hambrey *et al.* 1991) identified a clay interval interbedded with diatomaceous ooze. This was interpreted as representing deposition beneath an expanded Amery Ice Shelf, which, based on ^{14}C dates took place during the Holocene warm phase around 7000 a.

O’Brien (1992), interpreted several thousand kilometres of echosounder tracks and identified a number of sea-floor provinces based on the morphology of the sea floor. He described areas of intense iceberg turbation, sub-glacial forms, and rugged topography, and laterally extensive infills containing distinct internal reflectors.

DESCRIPTION AND DISTRIBUTION OF SEDIMENTS IN PRYDZ BAY

Sampling

The samples relevant to this chapter (Table 6.1, Figure 6.1) include material collected for this study and data and samples collected for previous studies, not all of which have been published. The data and samples include echosounder tracks of the seafloor, photographs of the seafloor of Prydz Bay, numerous short cores, grab samples, and pipe dredges from various studies, as well as two gravity cores.

<i>Data collected for this study</i>				
Year	Sample Type	Details		
1992/93	Cores	2 Gravity Cores, GC29, GC30		
	Grabs	32 Grab Samples		
<i>Data collected for previous studies</i>				
Year	Researchers	Sample type	Details	Publications
1982	Quilty, P.	Pipe Dredge	31	Quilty 1985
		Gravity Core	8	Quilty 1985
	Stagg, H.	3.5 kHz echosounder	8000 km ¹	Stagg 1985
1991	Franklin, D.	Short cores	15	Franklin 1993
		Trawl	3	Franklin 1993
		Grabs	1	Franklin 1993
Various	Quilty, P.	Bottom Photos ²	30	N/A

Table 6.1 – Data used for this study

¹ Of this only ~ 800 km are considered in this thesis.

² Bottom photos have been taken every summer season since 1991 with no reoccupations.

Sediment types encountered

The sediments encountered during this study (Figure 6.2) occur in two main provinces. In the south east of Prydz Bay there is an area dominated by fine grained muds, sandy muds, and sandy muds with dispersed dropstones. Around the edges of the bay, and extending along the shelf break along the Mac. Robertson shelf to the west are diamicts from clast poor muddy diamict to clast rich sandy diamict. There are a few places within this area where fine grained sediments are found, and these correspond to shallow parts of the shelf.

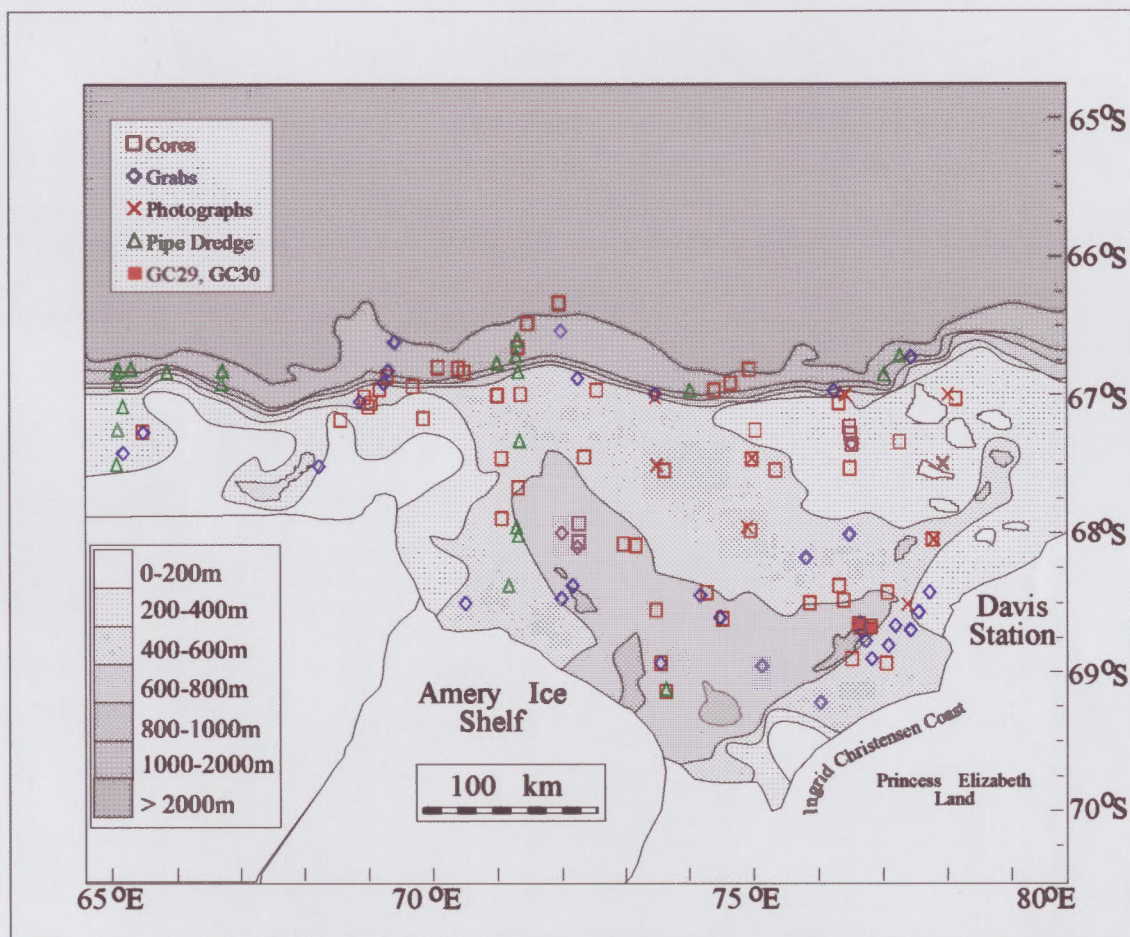


Figure 6.1 - Location of samples reported or referred.

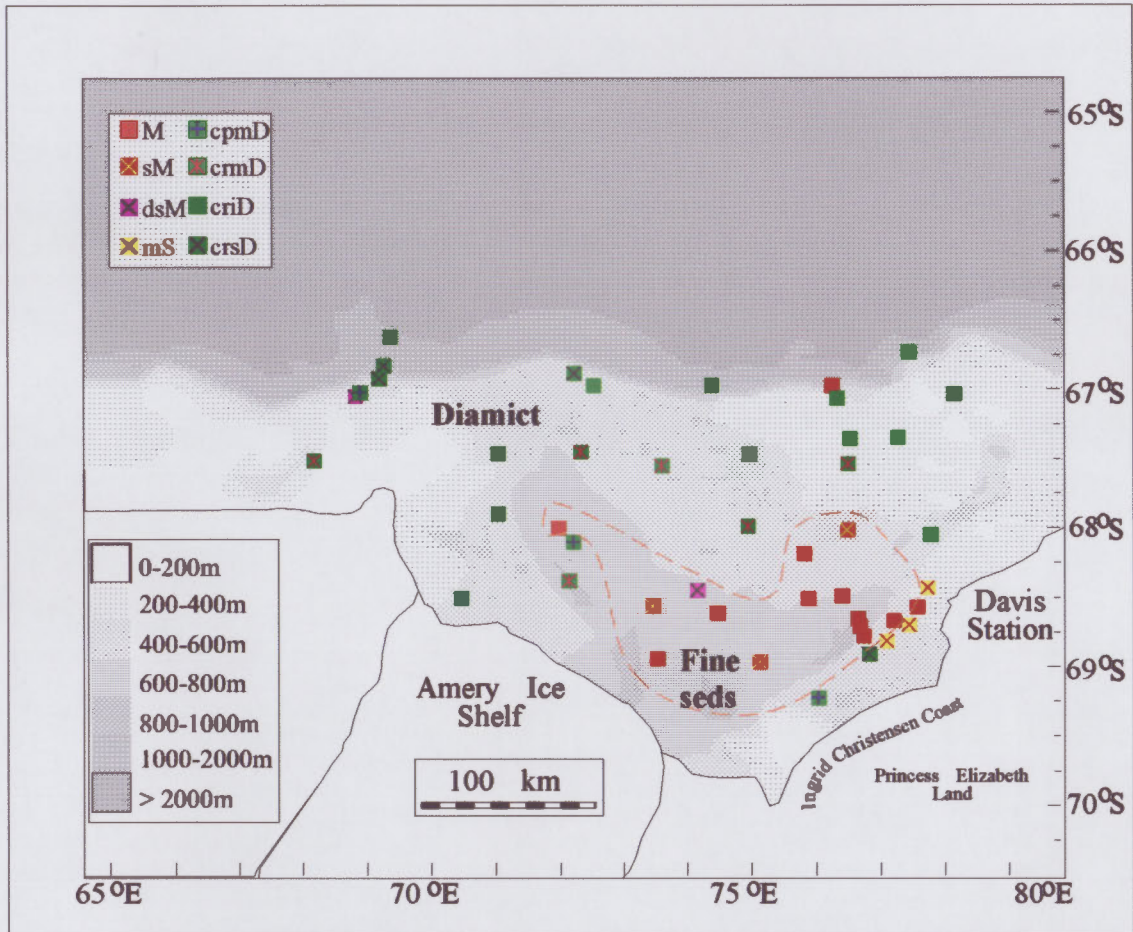


Figure 6.2 - Sediment distribution in Prydz Bay. There are two main provinces, a central area of fine grained sediments surrounded by diamict sediments. Classification after Hambrey (1994). D = Diamict, M = Mud, S = Sand, cp = clast poor, cr = clast rich, d = dispersed dropstones, i = intermediate, m = muddy, and s = sandy.

Bottom photos

All bottom photos are in Appendix E (Page 245). Fine grained sediments are found at 13 of the 14 sites (2, 5, 8, 9, 10, 11, 12, 13, and 14). Photographs of two of these sites (Figure 6.3) have a low diversity biota with the most common organism being an infaunal worm occupying a burrow and extending a feeding apparatus above the sediment. At these sites, the sediment surface is smooth but with circular mounds and depressions (~ 5 cm diameter) probably caused by sub-surface biological activity. The depressions appear to be filled with a drape of fine dark mud. In some instances, the worms are being pushed aside by bottom currents (Sites 1, 2, 9, 11, 12, and 13) and it is likely that these creatures are filter feeding on material brought with the current. There are surface feeding traces at some sites (Sites 8, 11, 12, and 14.) and a large (20 cm) species of annelid (?) worm (Site 2 & 8) may be responsible for some of those traces. Some traces are directly attributable to echinoderms (Sites, 11, 12, and 14). There is no apparent zonation of feeding traces with depth.

All of the sites at which fine grained sediments were photographed are within the area enclosing the fine grained samples collected with the exception of Site 5 which is located well in the north east of the bay (67.00°S, 76.49°E) on the northern flank of the Four Ladies Bank. The Four Ladies Bank may be protecting this area from both iceberg deposition and iceberg scouring.

The other sediment type is a coarser material found at Sites 1, 3, 4, 6, and 7 (Figure 6.4). At these sites the biota is much more diverse, often associated with an accumulating shelly hash made mostly up of the broken remains of bryozoa. Glass sponges are present.

This diamict sediment is typified by rounded to sub-rounded cobbles set in a sandy muddy matrix and with *in situ* encrusting epifaunal organisms. The sediment surface at Sites 1 and 3 is mostly horizontal but appears to dip steeply downwards across a very sharp boundary. This is interpreted as the edge of an iceberg scour although it may be the edge of a pile of boulders/cobbles that have been dumped and are currently in the process of being buried.

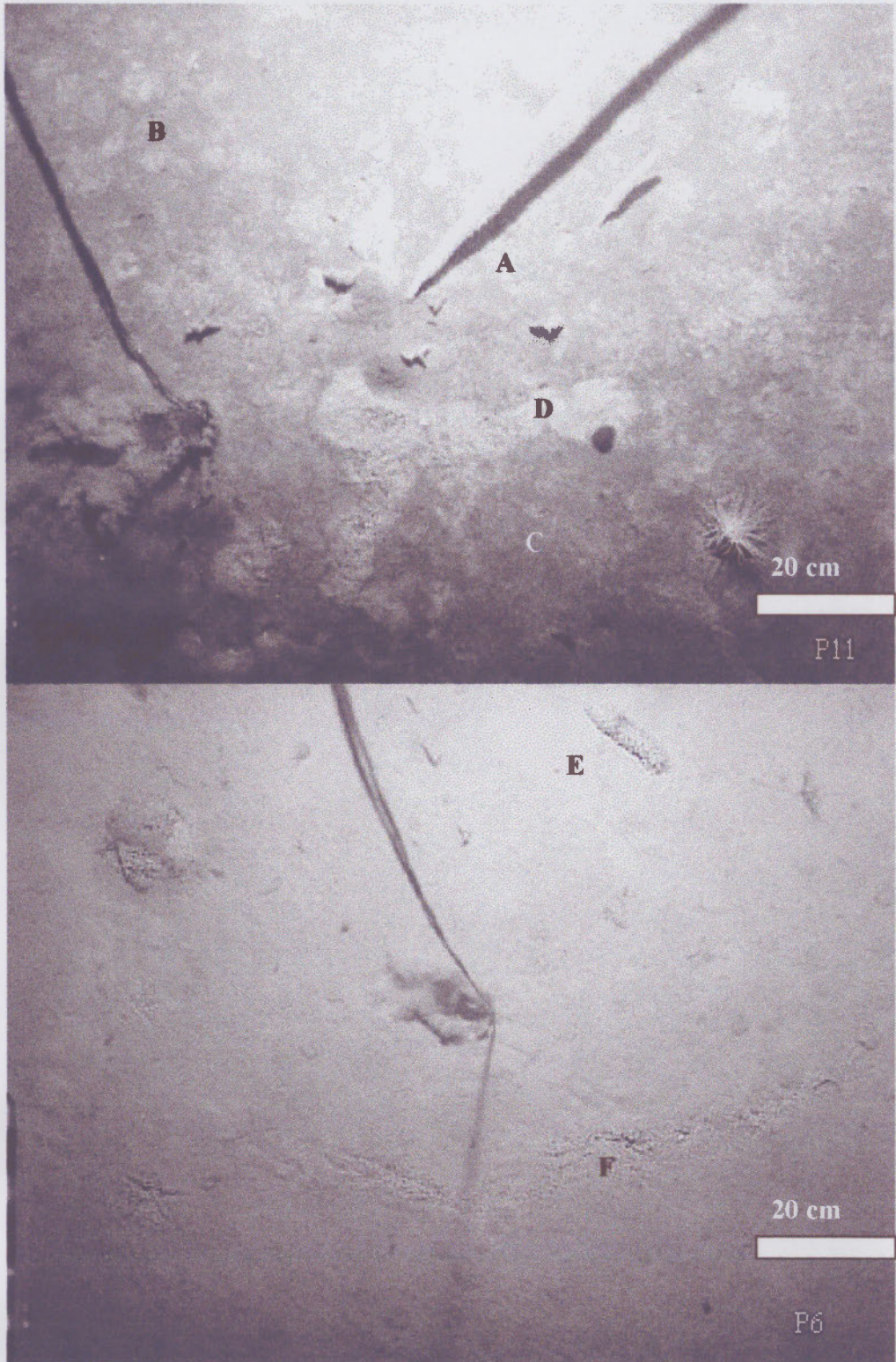


Figure 6.3 - Sea floor photographs of fine sediments showing an infaunal worm being displaced by the current (A), mounds and depressions (B) mud drape (C), echinoderm and its feeding trace (D), annelid worm (E), and unattributed feeding trace (F).

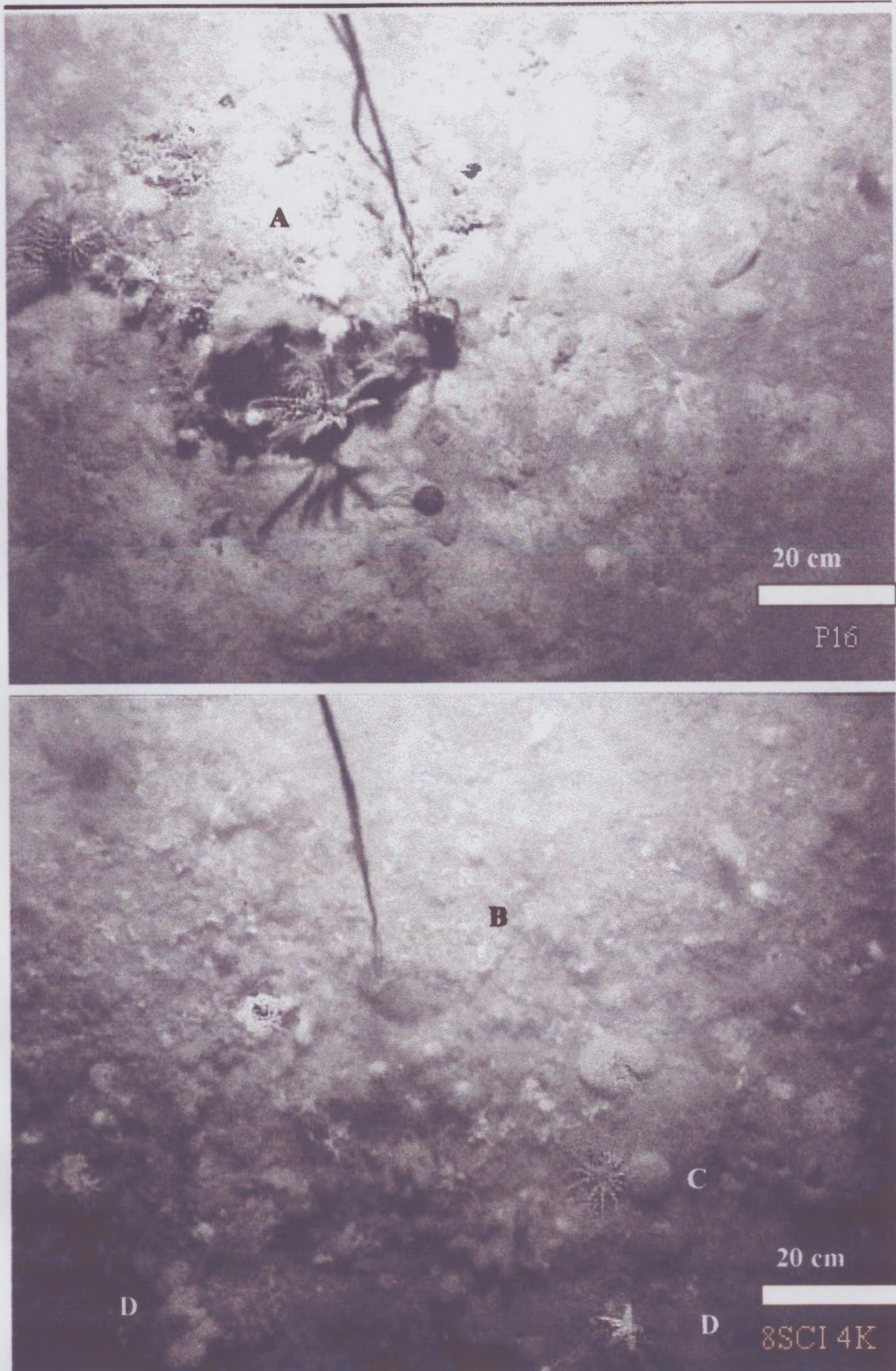


Figure 6.4 - Sea floor photographs of coarse sediments showing a large heavily encrusted dropstone (A), accumulating shelly hash(B), glass sponges (C), and a possible iceberg scour (D).

Of note are large dropstones encrusted with a large number of organisms at Sites 7, 10, & 14. This may indicate that the bottom waters contain sufficient oxygen and nutrients to support life but that the surrounding sediments probably do not. It may also be indicative of a slow rate of sedimentation but without details of the growth rates of the individual organisms no conclusions can be made in this regard.

Echosounder profiles

O'Brien (1992) identified a number of sea floor provinces based on the sea floor morphology observed in echosounder profiles. The echosounder traces for the area adjacent to the Ingrid Christensen Coast were re-interpreted and the following sea floor morphologies were observed.

Relict Till and basement

Relict till and basement seafloor (Figure 6.5), O'Brien's province 6, is typified by a very rugged appearance with chaotic reflectors and very high relief valleys and troughs, displaying many point reflectors suggesting individual large boulders. It is likely that this sort of seafloor represents the outcropping of the unit described by Stagg (1985) as PS 1. This sea floor type was encountered during the coring program on the KROCK voyage. It was typically a coarse, semi indurated and overconsolidated sediment which was found at the base of those cores collected adjacent to the Ingrid Christensen Coastline. It usually was represented only by a solid plug in the core cutter, the overcompacted sediment preventing further penetration of the corer. The overlying sediments consisted of both diamicts and muds.

Fine Grained Sediments

Fine grained sediments are widespread in Prydz Bay; in the Amery basin (Figure 6.6), and as infills to 40 m thick with internal horizontal reflectors - the most noteworthy such occurrence being in the Svenner Channel (Figure 6.7).

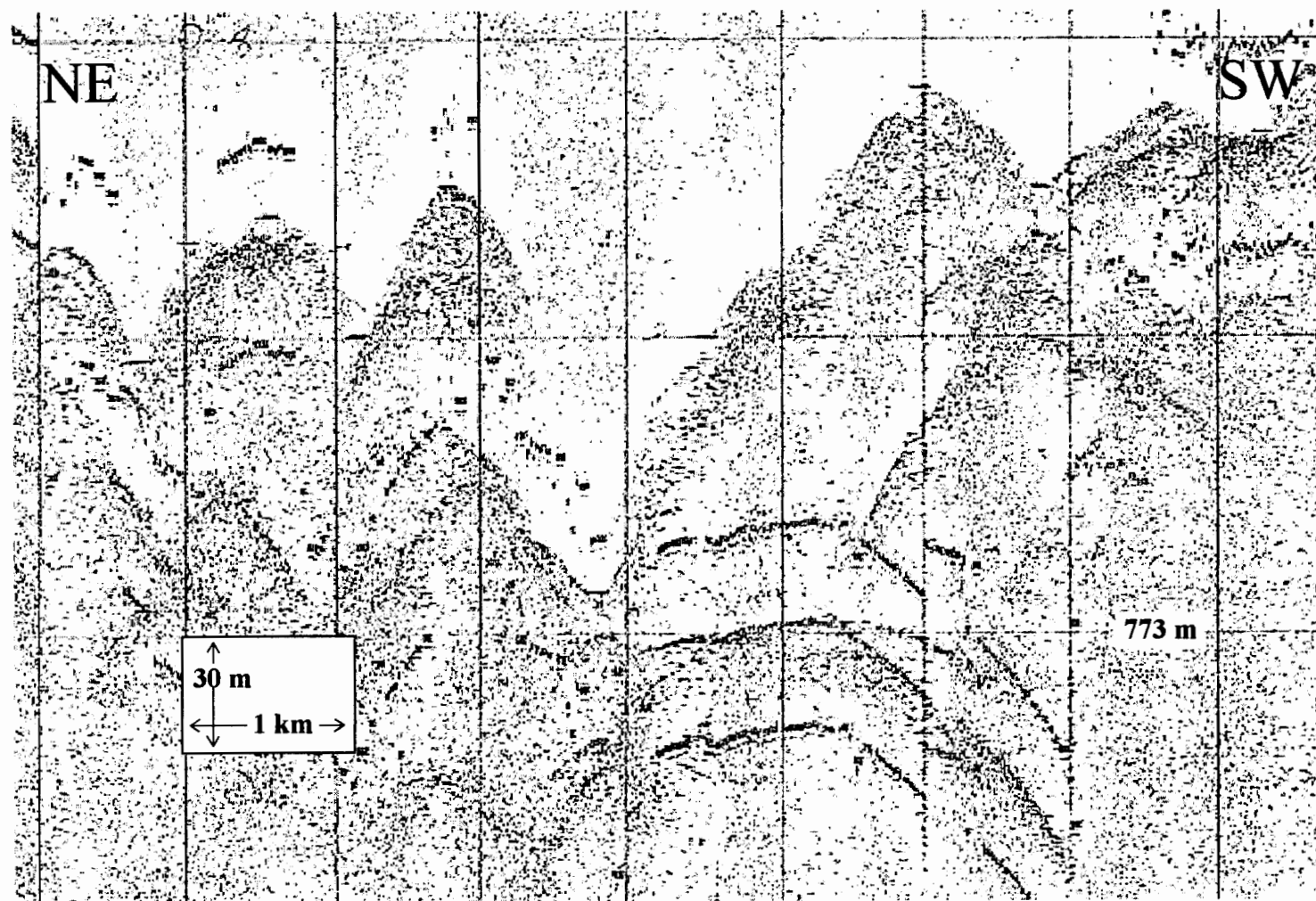


Figure 6.5 - Rugged sea floor underlain by relict till and basement rocks. O'Brien's Province 6.

The Amery Basin material, O'Brien's (1992) Province 2, has a very smooth surface with irregular ridges and swales, and drapes of fine sediments. The features in Figure 6.6, up to 10 m in vertical relief and several 100's of metres in width probably represent sub-glacial flutes or drumlins.

The Svenner Channel lens is also composed of fine grained sediment. It is laterally extensive and occurs in the bottom of the channel for much of its length. It may be deposited at this spot due to the topography of the channel creating a protected quiet water zone where decrease in current velocity results in fine material settling out of the water column. Gravity core KROCK GC 29 was collected from this deposit.

Iceberg Turbate

Iceberg turbate occurs extensively in Prydz Bay in depths between about 200 and 720 m. The profile of this seafloor is typically of low ridges and swales, but may include other features such as slumps and large scours. The ridge and swale topography is characteristic of fluted ice sheet topography and probably formed at the last glacial maximum beneath a more extensive grounded Lambert Glacier. The subsequent iceberg turbation results in the "jagged hashy" record from province 1 described by O'Brien (1992 fig 7.) O'Brien suggests that the unusually deep action of iceberg turbation is due to two factors. Firstly, the Amery Ice Shelf is up to 400 m thick at the calving line and so rolling tabular bergs may extend to depths of ~ 600 m. And secondly the 120 m depth further below this may represent the process occurring during a low sea level.

Distribution of seafloor morphologies

The sea floor types encountered (Figure 6.8) are distributed in three bands trending north east – south west. Adjacent to the Ingrid Christensen Coast, relict till and crystalline basement extend to the edge of the Svenner Channel which forms the boundary between the relict till and the fine grained sediments.

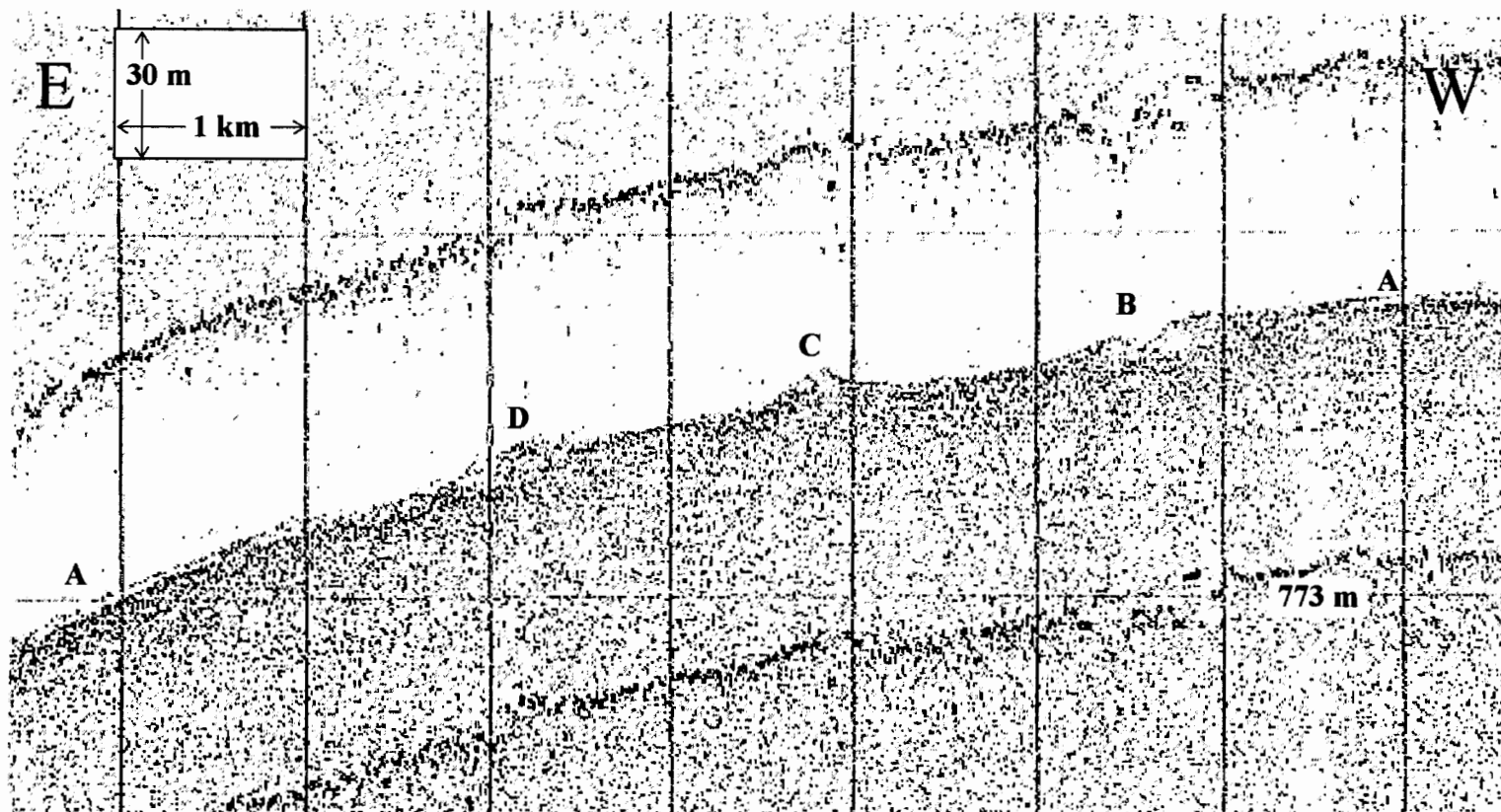


Figure 6.6 - Province 2 below the maximum depth of iceberg turbation. Drapes of fine sediment are seen at A. B, C, and D may be subglacial features, flutes or drumlins.

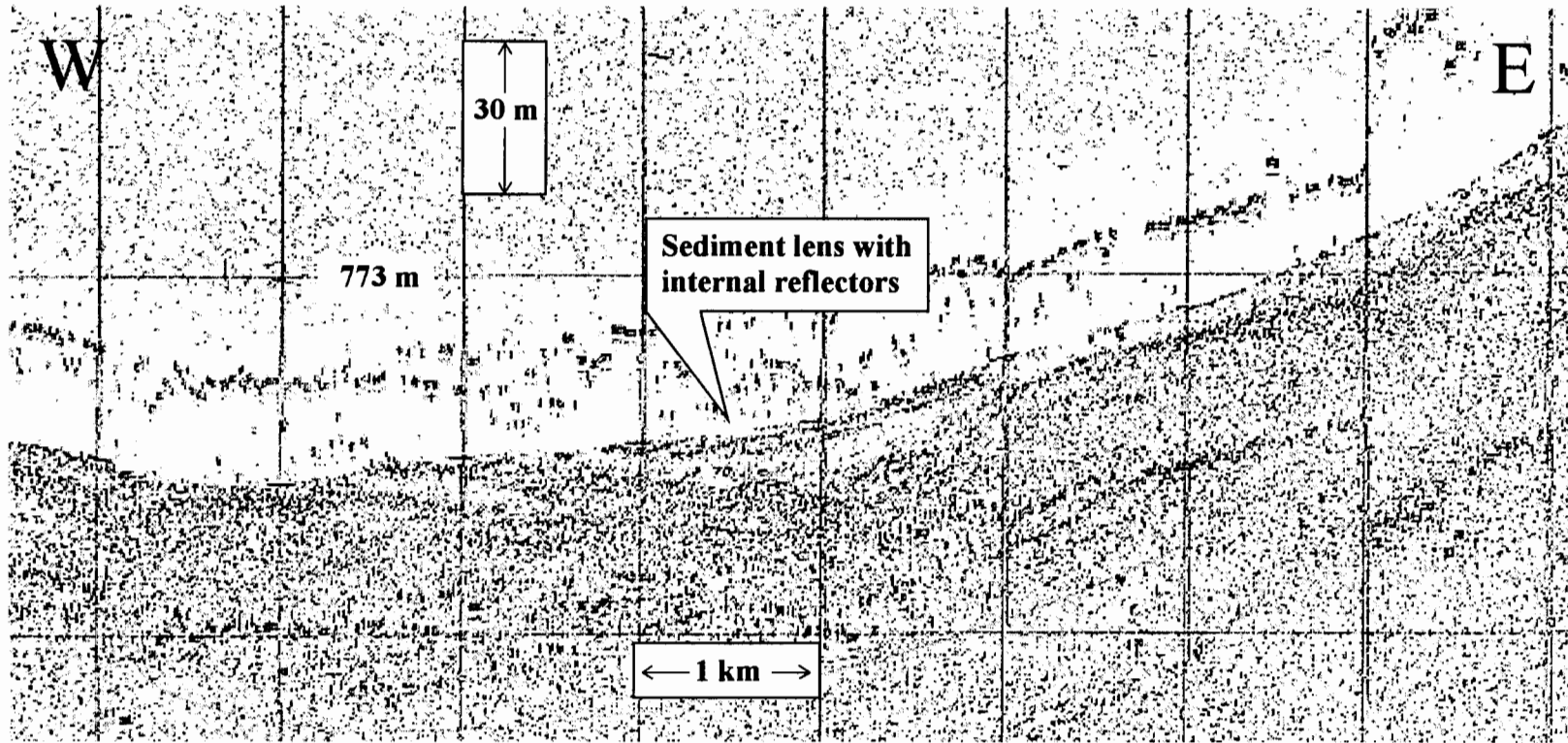


Figure 6.7 - Lens of fine sediment cored at GC 29. O'Brien's Province 5.

The stringer of fine sediments infills parts of the Svenner Channel and similar features appear to extend in an almost east west direction offshore from the Sørsdal Glacier and the Polar Record Glacier. O'Brien (1992) suggests that the Svenner Channel has been eroded by a more extensive Lambert Glacier the position of which was structurally constrained by the hard crystalline rocks at the edge of the Lambert Graben, and which preferentially eroded the softer riftogenic sediments within the graben.

Description of KROCK cores

Two cores were collected offshore approximately 40 km south west of the Vestfold Hills (Appendix F). Core GC 29 is a 352 cm long core consisting of a structureless clast-poor intermediate diamict at the bottom, fining upwards into a clast-poor muddy diamict which extends from approximately 320 cm dbsf to the core top. GC 30 is a more complex core of 150 cm length. The basal material is a clast-rich sandy diamict which extends to ~ 80 cm dbsf. There is an abrupt transition into a 15 cm thick sandy mud with convoluted laminations. This unit is overlain, again across a sharp boundary at 65 cm dbsf, into clast rich intermediate diamict which fines upwards through a clast-poor intermediate diamict then to a clast-poor muddy diamict at the surface.

While collected in close vicinity to each other the sedimentary processes and regimes are different. GC 29 was in the lens identified in the echosounder profile and GC 30 was collected in the sediment adjacent to the lens.

TOC ages for both of these cores indicate that, towards the bottom, both are ~ 25000 a old. GC 29 shows no sign of internal structure, erosional surfaces, lamination, or bioturbative structure, and appears to represent an uninterrupted record of sedimentation. The high levels of biogenic silica in this core suggest that, during its deposition, the area was subject to open marine conditions for most of the time. If this is so then the record of marine biogenic sedimentation suggests that the Amery Ice Shelf was not grounded in this region for the last 25000 a.

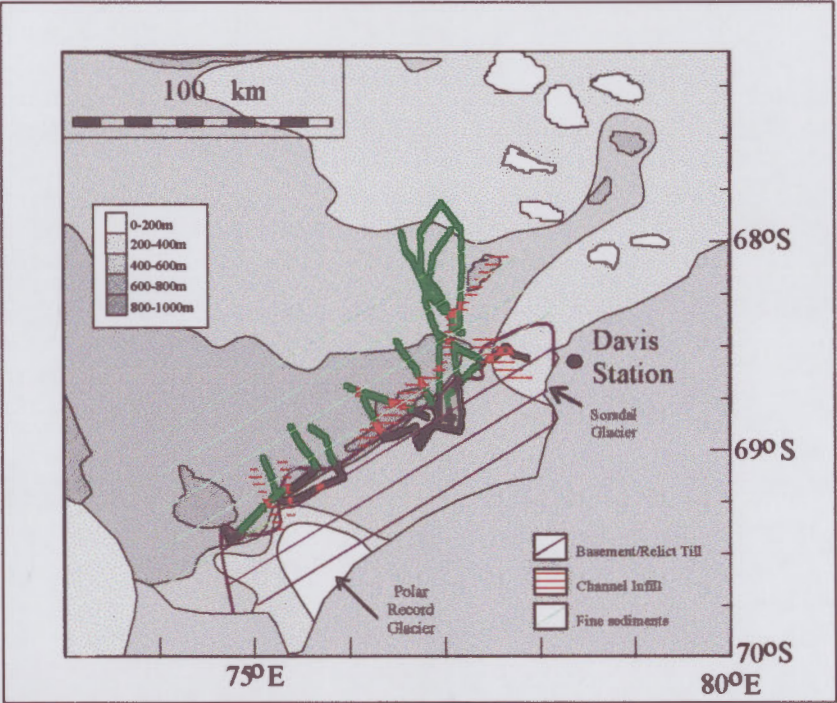


Figure 6.8 - Distribution of sediment types interpreted from seafloor morphology.

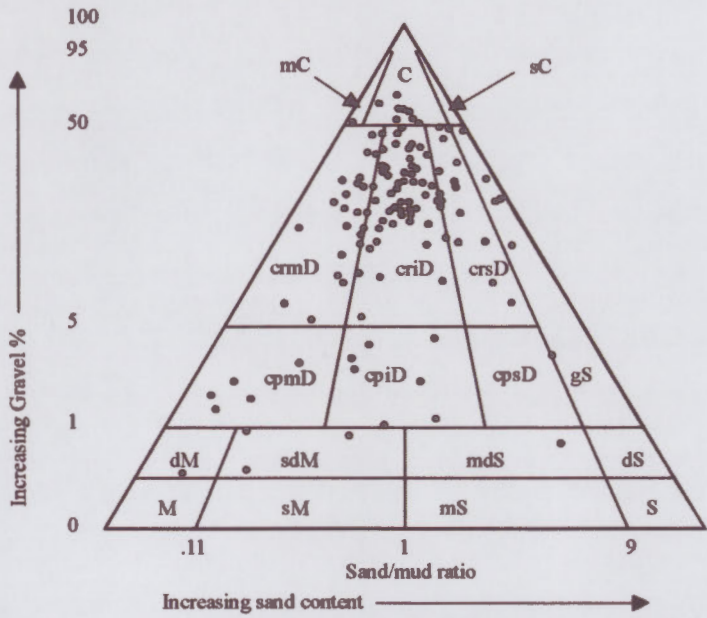


Figure 6.9 - Ternary plot of sediments sampled. Classification based on Hambrey (1994) but presented as a ternary plot. Note that the vertical axis is logarithmic to allow for the fine grained nature of the sediments. C = Conglomerate, D = Diamict, M = Mud, S = Sand, cp = clast poor, cr = clast rich, d = dispersed dropstones, g = gravelly, i = intermediate, m = muddy, s = sandy.

Viewed in the context of the depositional system in the Vestfold Hills over the last few tens of thousands of years where glacial ice does not appear to have extended as far as the coastline, it is possible that during the last glacial maximum between 10 and 25 ka (Adamson and Pickard 1984) the Lambert Glacier was not grounded near the Vestfold Hills. These data suggest a significant inlet of marine water along the Ingrid Christensen Coastline during the last glacial maximum and support evidence that the east and west sides of the Lambert Glacier responded to climate changes differently (O'Brien 1992, Domack *et al.* In Prep). Domack *et al.* (In prep) suggest that the Lambert Glacier did not reach the shelf edge in the last glacial maximum, but in the maximum before the last.

Sediment parameters

Texturally, Prydz Bay surficial sediments form a continuum from very fine grained sediments composed of biosiliceous remains of phytoplankton and fine grained terrigenous material to coarse grained diamicts and conglomerates (Figure 6.9). The separate fractions of these sediments show no variability with depth (Figures 6.10, 6.11, & 6.12)

In most environments, the deposition of fine grained material is indicative of low energy levels and the deposition of fines generally increases with increasing depth as current action at depth is usually much less than in shallow areas. These data suggest that current action is not an important determinant in the deposition of fines. Nonetheless, particular features do appear to be related to current regime, in particular the lens of fine grained material discussed earlier.

The plot of mean grain size versus standard deviation for the sand fraction of Prydz Bay sediments (Figure 6.13) demonstrates that there is no recognisable grouping within the sediments in these parameters, regardless of the fact that the sand fraction is present in most samples from very fine grained sandy muds to coarse grained conglomerates. This leads to the conclusion that the process supplying these sands to these locations is ubiquitous throughout Prydz Bay.

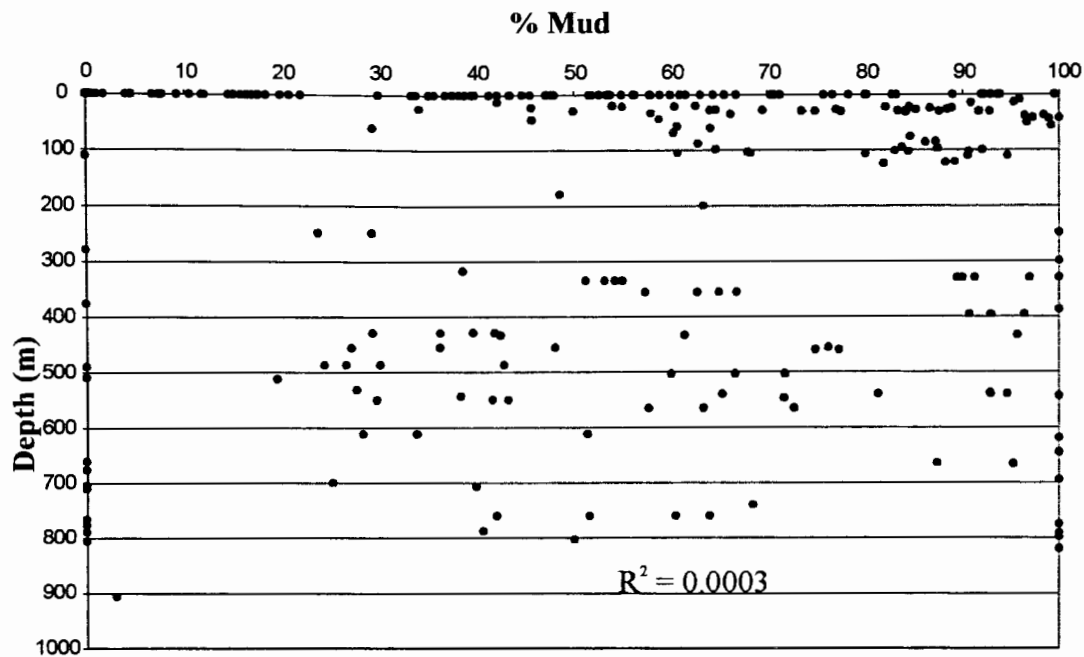


Figure 6.10 - Bivariate plot of mud against water depth which indicates that processes depositing the mud fraction are independent of depth.

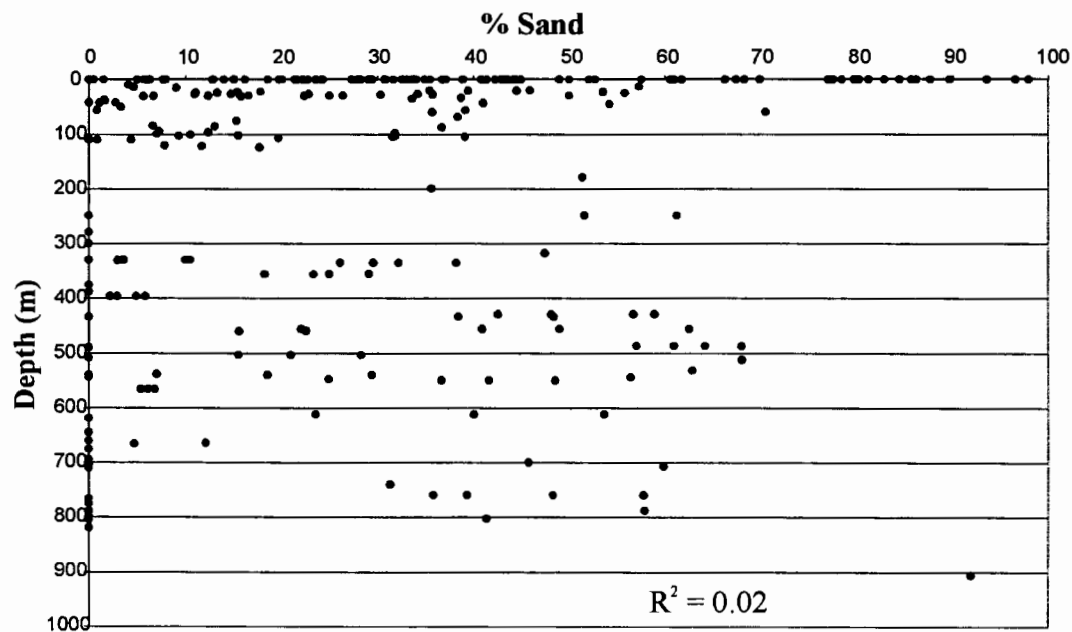


Figure 6.11 - Bivariate plot of sand against water depth which indicates that processes depositing the sand fraction are independent of depth.

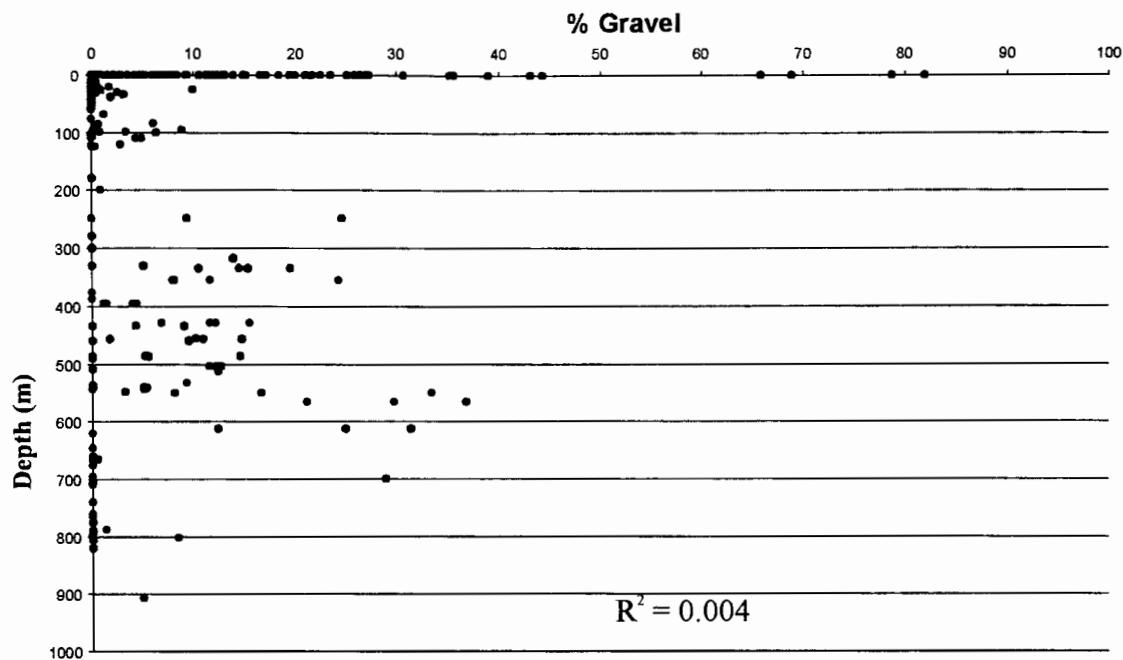


Figure 6.12 - Bivariate plot of gravel against water depth which indicates that processes depositing the gravel fraction are independent of depth.

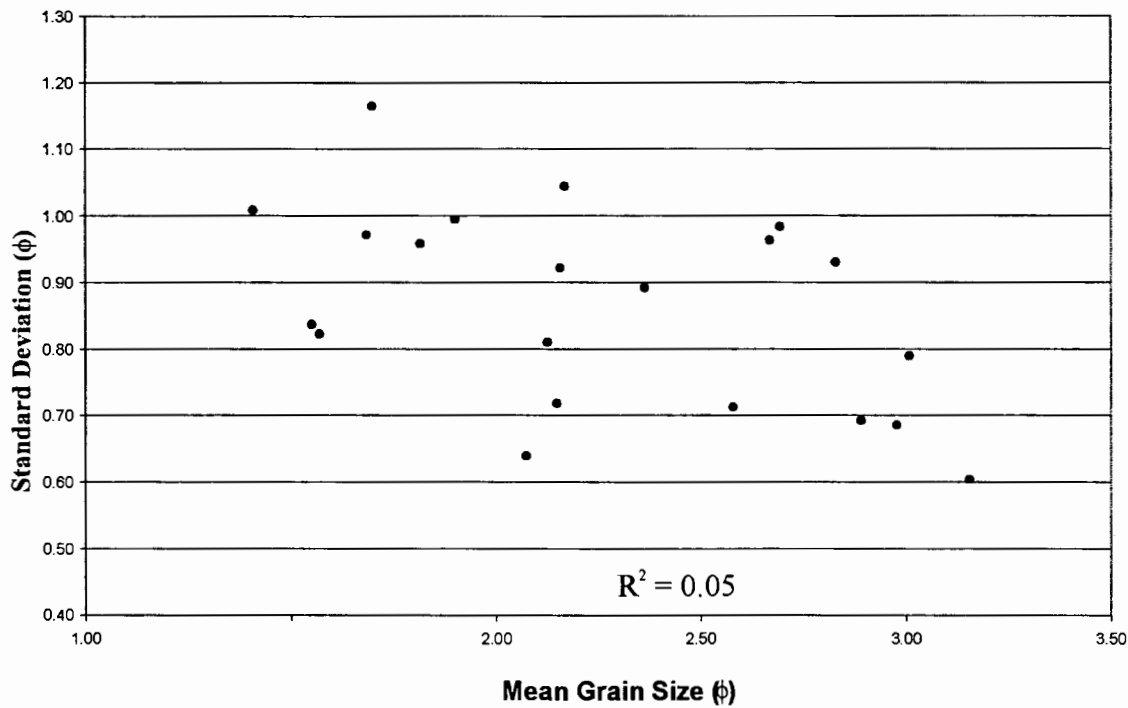


Figure 6.13 - Bivariate plot of Mean Grain Size versus Standard Deviation

PROCESSES

Biogenic deposition

Biogenic deposition occurs in two separate processes: firstly the production and accumulation of, biogenic siliceous particles and, secondly, calcareous skeletal remains. Of these the more important is that of the siliceous organisms which make up a significant proportion of all of the sediments in Prydz Bay. In contrast, the accumulation of calcareous organisms is generally less important but with significant accumulations of biogenic carbonate occurring in isolated areas to the north of Prydz Bay.

Silica

Biogenic silica is high in offshore sediments due to the predominance of diatom production, with values as high as 75% biogenic Si by mass. There is no apparent variation in the proportion of biogenic silica with depth (Figure 6.14) suggesting that the deposition of biogenic silica is not depth controlled.

The siliceous component of the sediments of Prydz Bay is composed of the skeletal remains of a variety of organisms, mainly diatoms (Taylor *et al.* In press), but also including other groups such as Parmales (Franklin and Marchant 1995). The relative ease of study and reported dominance of diatoms over other planktonic groups in the Southern Ocean (Fenner *et al.* 1976, Defelice and Wise 1981 and references cited.) have ensured that most studies have focussed on them.

Diatom biomass is high with counts in Prydz Bay being as high as $6 \times 10^8 \text{ l}^{-1}$ (Koslova 1970). Diatoms bloom in the waters around Antarctica in the spring at the sea ice edge (Smith and Nelson 1986b). Their presence after the dark winter is probably due to inoculation of the water by cells that were incorporated into the sea ice in the previous summer (Garrison and Buck 1989). Diatom blooms follow the ice edge as it retreats towards the Antarctic continent during spring.

Phytoplankton communities in most aquatic environments undergo a seasonal succession of species in response to changing light climate, temperature or nutrient conditions (McMinn and Hodgson 1993, Smayda 1980, Reynolds 1984). Perrin *et al.* (1987) noted a marked seasonality in phytoplankton abundance and species composition at a coastal marine site near Davis Station in 1982. While measurements in Prydz Bay have been concentrated in coastal areas, this pattern of bloom and bust probably occurs unevenly throughout the study area in any particular year but it is likely that all areas of the bay are equally affected by diatom production over time.

Siliceous remains settle through the water column until they reach the sea floor as individual particles or as a component of marine snow. The time that it takes for individual particles to settle to the seafloor may result in the decoupling of the seafloor sediments from the planktonic productivity immediately above as a result of current action which translocates settling particles. The settling process, however, mainly occurs via predation. Diatoms reaching the sediment via faecal pellets display a very poor preservation (Baldauf and Barron 1991). Franklin (1993) reported that >80% of diatom frustules from Prydz Bay were badly broken however, Dunbar *et al.* (1991) reported that nearly all biogenic opal delivered to the mid and deep water column consisted of unbroken diatom tests of the genus *Nitzschia*, many of which settled within faecal pellets. Gersonde and Wefer (1987) estimate that up to 50% of biogenic silica particles in Bransfield Strait are incorporated in the faecal pellets of zooplankton such as *Euphausia superba* (Antarctic krill), and *E. crystallorophias* (Crystal Krill). Faecal pellets are significantly larger than the remains of the individual organisms predated, settle through the water column quickly and are much less likely to be reworked by bottom currents. It is possible, however, that once deposited, biogenic remains could be remobilised by bottom currents, especially if bioturbation and infaunal bacterial productivity disaggregate the faecal pellets.

A second source of biogenic silica is from benthic siliceous organisms, in particular, glass sponges (Figure 6.4). These sponges have a dense skeleton of

long spicules which fall to the seafloor on the death of the sponge. These organisms mainly occur in sandy or gravelly sediments. The proportion of the sediment represented by this material is difficult to establish but is significant in some places.

The majority of diatoms fall within the fines fraction ($< 63 \mu\text{m}$), and this is reflected in a relationship between the proportion of biogenic silica and the proportion of mud with Si increasing with increasing mud proportion (Figure 6.15).

Within Prydz Bay, biogenic silica is widespread (Figure 6.16) but subject to lateral variability. Most low silica samples are found on the Four Ladies Bank, along the shelf break, and along the western side of the Prydz Channel. This pattern probably reflects the action of bottom water currents (to be discussed later) which prevent the settling of fines in all but protected areas such as old iceberg furrows. In the areas dominated by fine grained sediments, biogenic silica forms up to 75% of the sediments. In the areas where coarser sediments, diamicts, are found, biogenic silica is present in lower proportions.

As it is likely that biogenic silica is being produced in the surface waters over all of Prydz Bay, and that such production is probably evenly spread over time, then the differences in bio-silica content of the sediments must be due to other processes including oceanographic which remobilise fine grained material and move it to depositional basins, the dilution of biogenic silica by the deposition of terrigenous material. As a proportion of the whole, biogenic silica comprises 8.61% of the surficial sediments of Prydz Bay.

CaCO₃

CaCO₃ is present in many Prydz Bay sediments (Figure 6.17) and samples fall in two broad categories. The first consists of sediments with high percentages of CaCO₃ found in samples from the shelf break in the north west of the bay with one sample (S111) collected at a depth of 396 m and having ~ 50% CaCO₃ by mass.

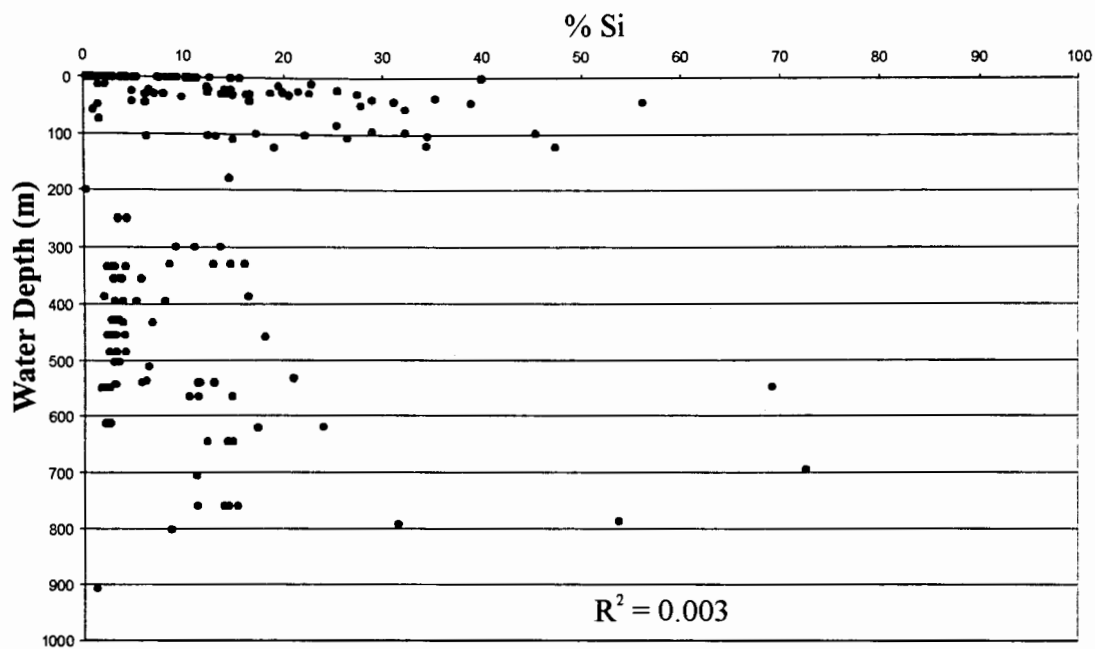


Figure 6.14 - Bivariate plot of biogenic Si against water depth which indicates that processes depositing the silica are independent of depth.

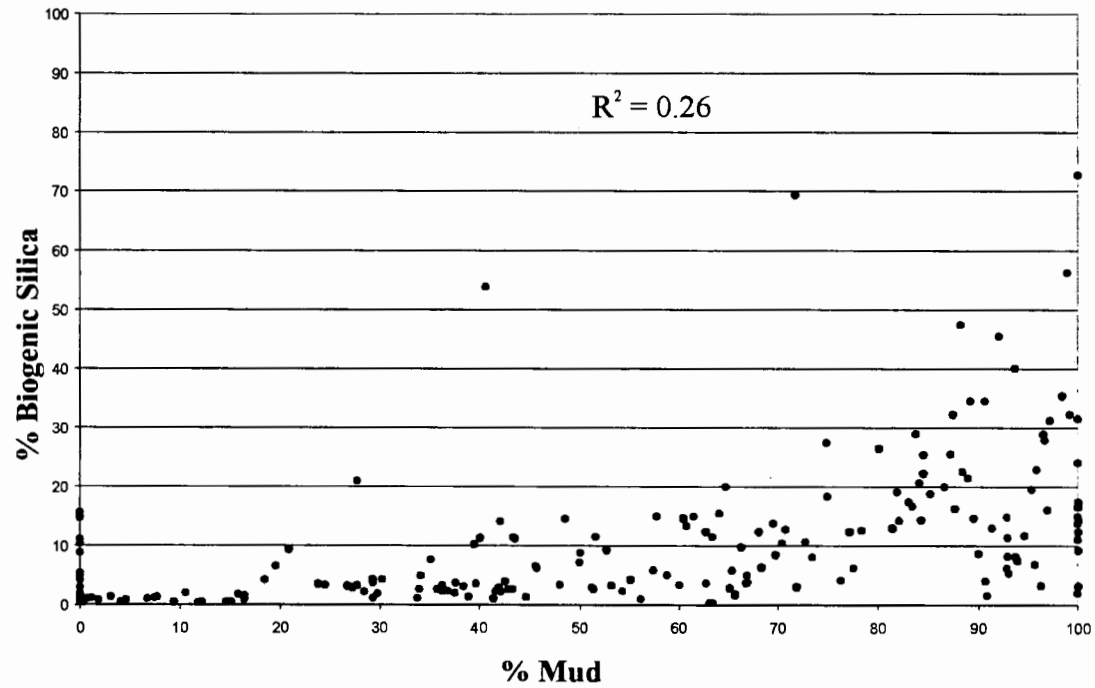


Figure 6.15 - Bivariate plot of biogenic silica and %mud shows a loose association between the proportion of fine grained sediment and biogenic silica.

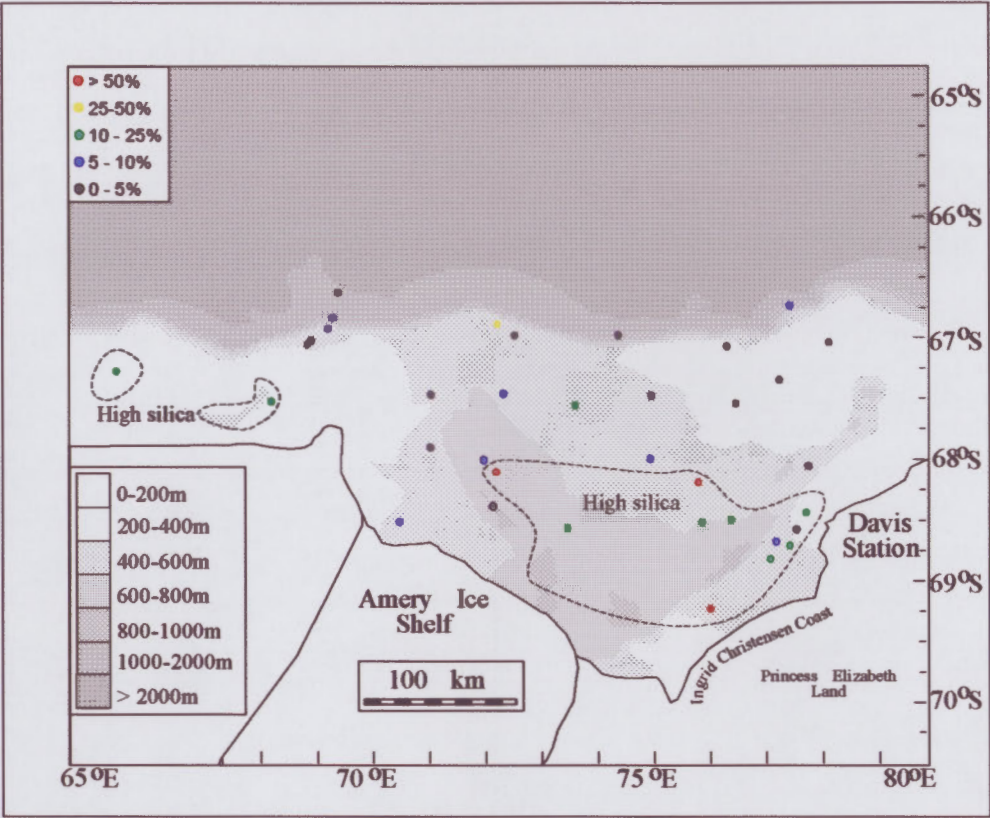


Figure 6.16 - Biogenic Silica in Prydz Bay sediments

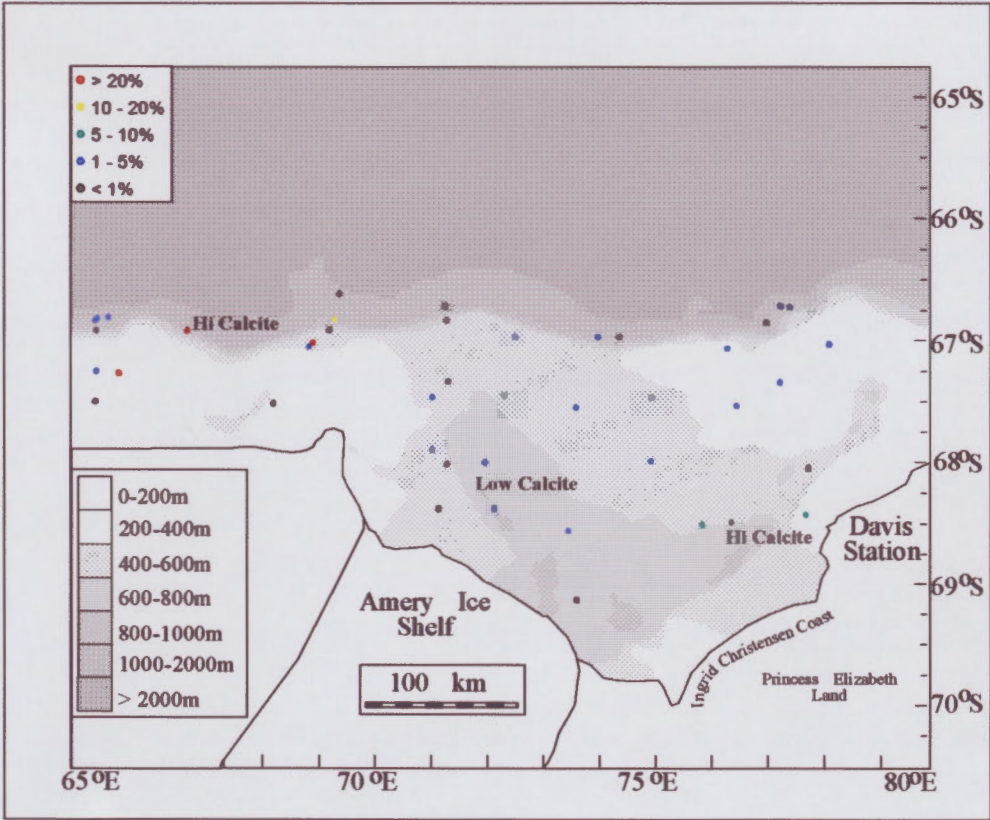


Figure 6.17 - Biogenic Calcite in Prydz Bay sediments

Two other samples collected from below 800 m, GR27 and PQ13, have high values of 17.8 and 3.3% by mass respectively. These samples were collected to the north of Prydz Bay and represent the more open marine conditions. Other high values are found in the south east of the bay near to the coast. The second grouping consists of sediments which are low in CaCO_3 and can be found at all depths while CaCO_3 rich sediments are restricted to depths above 800 m.

CaCO_3 varies with depth (Figure 6.18) with CaCO_3 values reaching a maximum at approximately 350 to 400 m and then decreasing to zero at around 800 m. These data suggest that the Calcite Compensation Depth (CCD) is shallower within Prydz Bay (at about 800m depth) than the deeper CCD found around the rest of the Antarctic continent (~ 1500 m). In sediments shallower than 800 m calcite averages approximately 3.5% by mass. Assuming that the data set accurately represents all depths and sediment types, the average percentage of CaCO_3 by mass for all sediments in Prydz Bay is 4.83%.

The deposition of biogenic CaCO_3 is mainly due to the *in situ* growth of organisms producing calcareous or aragonitic skeletons. The calcareous organisms found in Prydz Bay form a diverse and disparate assemblage, including foraminifera, pteropods, brachiopods, bivalve molluscs, ostracods, gastropods, and echinoderms, among others. In particular, one site ($66^\circ 48.59'S$ $70^\circ 23.52'E$ – water depth 880 m) yielded a sediment composed almost exclusively of the disassociated plates of barnacles (Figure 6.19), with some living and abundant brittle stars. The majority of the work conducted on calcareous fauna has, however, been on the foraminiferids.

While some planktonic foraminifera have been identified in Prydz Bay sediments, mainly at sites in the north of Prydz Bay where there is the most marine influence (Quilty 1985, Franklin 1991 & 1993), they are restricted to one species, *Neogloboquadrina pachyderma* (Ehrenberg). This planktonic species forms a minor part of the foraminiferal associations with the majority being benthic forms (Franklin 1991&1993).

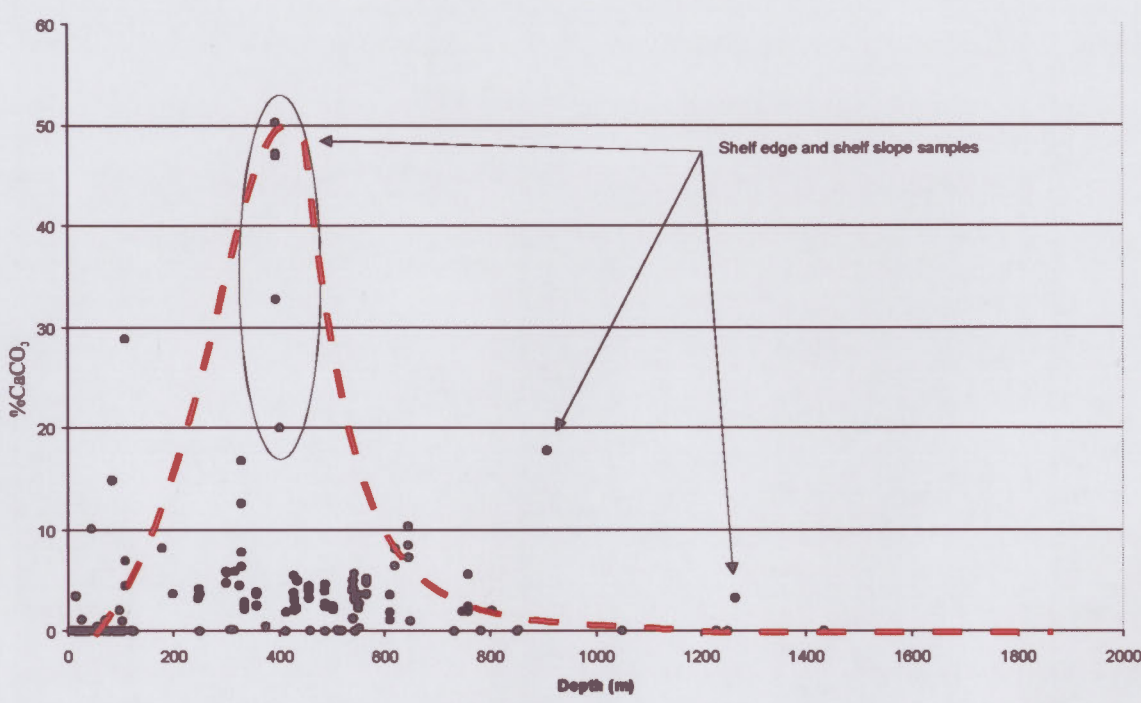


Figure 6.18 - Bivariate plot of calcite against depth shows an optimum depth of calcite preservation at ~ 350 m and absence of calcite in depths below 800 m.

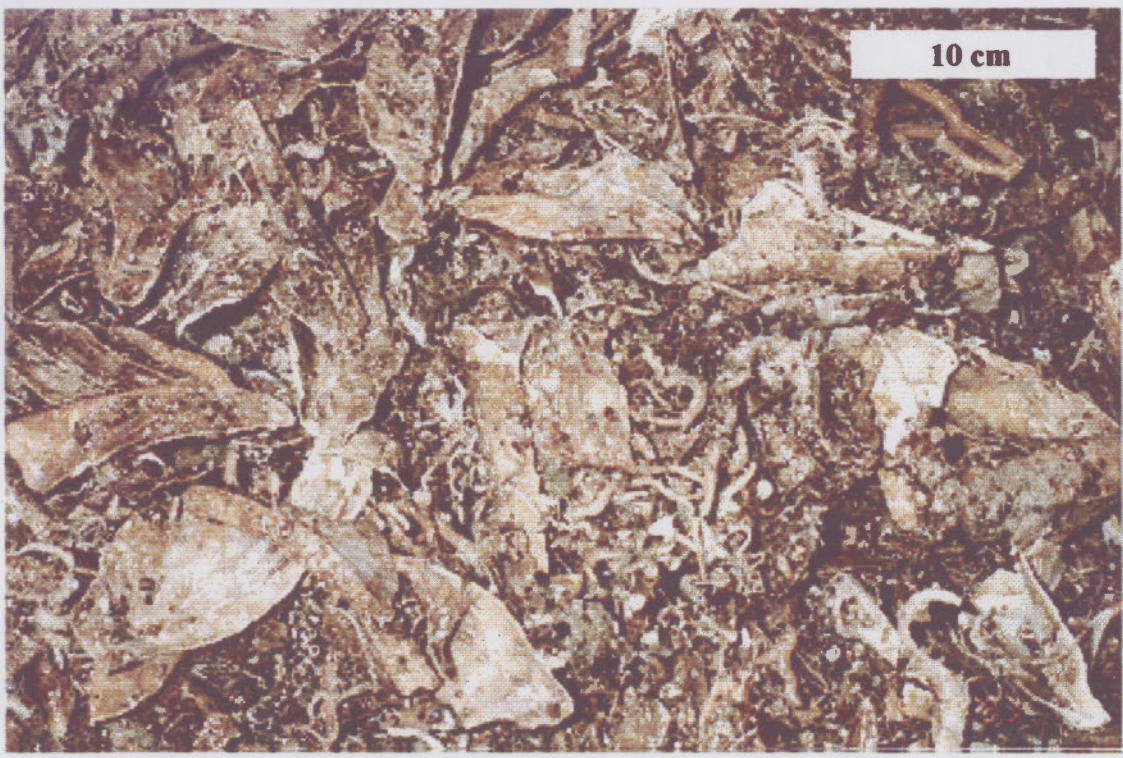


Figure 6.19 - Carbonate rich sediment composed of barnacle plates and ophiuroids. (Near GC 13 - 66 48.59 S, 70 23.52 E, 880 m)

Benthic foraminifera are found throughout Prydz Bay in low numbers. Faunas have low diversity. Quilty (1985) demonstrated that the benthic foraminiferal faunas were dominated by agglutinated forms and Franklin (1991 & 1993) found that the calcareous forms were restricted to *Triloculina rotunda* d'Orbigny and *Triloculina trigonula* (Lamarck). The low diversity of calcareous foraminifera in Prydz Bay is due to the shallow Calcium Compensation Depth (CCD) discussed later. The two *Triloculina* species, being infaunal, are able to exist below the CCD by manipulating the pore waters in the bottom sediments to ensure they are saturated with respect to calcite. Preservation of all forams is good, with even infaunal benthic forams living below the CCD being well preserved (Franklin 1991).

Quilty (1985) identified a north-south trending line roughly corresponding to 72° E to the east of which faunas are dominated by calcareous foraminifera and to the west of which agglutinated forams dominate. This finding is likely to be, at least to some degree, affected by the sampling pattern, and the distribution of calcareous versus agglutinated forams is more complex.

The Carbonate Compensation Depth

Based on the occurrences of calcareous foraminiferal faunas, Quilty (1985) concluded that the CCD in the areas to the north of Prydz Bay was between 1220 and 1500 - 1600 m in depth, and commented that this was at variance with other studies (Osterman and Kellogg 1979) which found evidence of the CCD being between 400 to 700 m in the Ross Sea. The CCD on the George V continental shelf and slope has been described as multibathic (Anderson 1975, Milam and Anderson 1981) occurring at depths as shallow as 336 m and as deep as 587 m based on the distribution of a transitional shelf foraminiferal assemblage, however, it is difficult to compare that study with this or Quilty's due to differences in methodology. Quilty (1985) further suggested that the CCD is not the controlling influence on foraminiferal distribution and that oceanographic factors such as nutrient control must be invoked to explain the observed distribution.

The variation of calcite with depth in Prydz Bay (Figure 6.18) shows two separate systems. The first represents the samples from within Prydz Bay. The percentage of carbonate drops dramatically from a high value of 50% at ~ 420 m to values in the order of 5 - 10% at depths greater than 600 m. Thereafter, the values decrease steadily to 0 at ~ 820 m. Thus the CCD in Prydz Bay is in the vicinity of 820 m while the Lysocline seems to occur at a depth in the vicinity of 450 m. The high values in the diamict samples from the north of the shelf break are in the area in which Quilty (1985) suggested the CCD was deeper than 1200 m. There the CCD is in the vicinity of 1300 – 1400 m.

Another process that may affect the amount of calcite in sediments is the rafting of calcite from other areas. Calcite could be rafted from other areas via sea-ice freeze from shallow coastal areas. However, the effect is of little importance in considerations of CCD as the major concern is preservation of calcite. In this instance, rafted calcite can give as much an indication of the presence of a CCD as *in situ* calcite.

Terrigenous Deposition

Ice-rafting

Iceberg meltout is the process whereby englacial debris entrained in glacial ice which is melting into the surrounding water, is released to settle to the sea floor below. The rate at which sediment is released is directly affected by the rate at which the iceberg melts. Melting occurs in two main ways, the basal melting of ice in direct contact with seawater and melting via insolation of the area of an iceberg exposed to the atmosphere.

Basal melting is dependent on the temperature differential between the glacial ice and the sea. Russell-Head (1980), working on small ice blocks floating in saline water of different temperatures, used the following equation;

$$M_b = 2.08 \times 10^{-7} (T_s + 1.8)^{1.5} \quad \text{Equation 6.1}$$

where M_b is the basal melt rate in ms^{-1} and T_s is the sea water temperature ($^{\circ}\text{C}$). Weeks and Campbell (1973), based on theoretical considerations, included some external parameters when arriving at

$$M_b = 6.74 \times 10^{-6} v_w^{0.8} \frac{\Delta T}{X^{0.2}} \quad \text{Equation 6.2}$$

where v_w is free stream relative horizontal water velocity past an iceberg, ΔT is the temperature differential between an iceberg and the surrounding water and X is the length of an iceberg side. Russell-Head (1980) established that melt rate was higher when the temperature of the water was higher and Weeks and Campbell (1973) indicated that the higher the temperature differential and the water flow rate, the higher the iceberg melt rate. Weeks and Campbell (1973) stated that the melt rate is relatively insensitive to the size of the iceberg, but small bergs have a higher melt rate than large bergs. An important implication of these equations is that where the temperature differences are very low or zero, then the basal melt rate is zero or close to zero. As the temperature of the waters of Prydz Bay are below the melting point of fresh ice (-1.5°C) for all but the height of summer (Quilty 1985), it is likely that meltout from icebergs is low.

Iceberg meltout sediments can be recognised in the fossil record by the large range of grain sizes. In practice, however, it is probably not easy to distinguish between iceberg meltout sediments and sediments from other ice rafting processes, ie. sea ice rafting and iceberg rollover and dump.

Ice dumping

As icebergs undergo basal melting they become less stable and periodically overturn thus exposing basal ice rich in englacial debris to the atmosphere. As heat from the sun melts exposed ice, englacial debris is concentrated on the surface of the iceberg because the dark coloured sediment decreases the albedo and more heat is absorbed. Østrem (1959) empirically derived sub-aerial meltout rates and established that the rate is dependent on the thickness of the

sediment layer on the ice surface, with surface debris cover up to 0.5 cm thick enhancing the melting process. At greater thicknesses the melt rate is progressively retarded. Eventually, some of the accumulated sediment slumps and flows off the side of the iceberg (Clark and Hanson 1983). With time the iceberg overturns and the sediment which has accumulated on the top of the berg is dumped in a single event. As the basal meltout rate for icebergs within Prydz Bay is very likely to be minimal due to the generally cold waters, it is probable that sedimentation via iceberg dumping is a more important process than iceberg meltout.

The erratic nature of these individual sedimentary events and the restricted nature of the subsequent sedimentary deposits on the sea floor is probably governed by the erratic nature of iceberg overturn events (Dowdeswell and Murray 1990). Over time, however, the random iceberg rollovers probably result in dump sites spread over wide areas of the floor of Prydz Bay, with the distribution of the resultant sediment dependent on the drift tracks of icebergs through the area. Iceberg drift tracks have not been studied in Prydz Bay, but a little is known about the major ocean currents in the bay and they will be discussed later.

These glacially derived and ice rafted sedimentary deposits can be recognised by the presence of all size fractions; mud, sand, and gravel. As stated, there is probably limited opportunity to distinguish between iceberg meltout and iceberg rollover derived sediments in any but the largest scale where iceberg rollover sediments will exhibit local aggregations of very coarse material. In some instances, if the dumped sediment has fallen through a strong current, the finer fractions will spread out in a "tail" trending downstream. These sedimentary structures would be visible in outcrop but very difficult to recognise using marine sampling or imaging techniques.

Sea-ice rafting

Finally, debris can be rafted by sea-ice. As discussed in Chapter 4, aeolian transport of quite coarse material occurs in the Vestfold Hills. Debris is transported onto the sea-ice during the winter and is incorporated in the sea-ice by being trapped by snow which falls on top. In the spring, the higher albedo of the dark debris allows it to move downwards through the ice because of differential heating of the darker particles. When the sea-ice breaks out in the summer, it is transported by the east wind drift and surface ocean currents, and releases debris into the water column as it melts.

Areas of high gravel values can be used to give an indication of where these processes are occurring (Figure 6.20). The plot of gravel percentage shows high gravel values on the shelf and shelf break in the north west of the bay and in an area which rings the deeper parts of Prydz Bay enclosing a zone with no gravel content. This distribution may be the result of a number of factors. Gravel content may be increased (above that supplied by ice rafting) by iceberg turbation which mixes underlying subglacial sediments with material currently being deposited. Secondly, the degree of ice rafting will be determined by the iceberg drift patterns which are in turn dependent on the oceanographic regime in Prydz Bay. Both of these factors will be discussed later. In general, however, the patterns of gravel content point to ice-rafting occurring around the margins of Prydz Bay, probably a result of ice rafting by icebergs and sea-ice in the east wind drift and the Prydz Bay Cyclonic Gyre which are discussed later.

Whatever the mechanisms depositing terrigenous sediments, the three fractions, gravel, sand and mud comprise 6.20%, 23.60%, and 56.77% of the surficial sediments respectively.

Reworking processes

Bioturbation

Bioturbation is the process by which organisms which reside in the sediment mix those sediments in the course of their living processes. Infaunal organisms churn the sediments as they move through them or physically move sediments during the construction of burrows. Some organisms ingest the sediment, extract nutrients and excrete the remainder. This process results in the homogenisation of the top 10 to 15 cm of sediment. Some organisms excavate burrows to much greater depths and some molluscs, for instance, produce large burrows which can be infilled by sediments much newer than the sediments in the burrow walls. Thus, in areas of a sedimentation rate of 0.1 mm a^{-1} , infills of 20 cm deep burrows may place new sediments adjacent to sediments up to 2000 years old. This situation compromises the accurate dating of marine sediments.

Bioturbation can be recognised at most scales of observation. The activities of infaunal and epifaunal organisms leave feeding traces and burrow infills which can be recognised in cores and on depositional surfaces.

Bottom photographs indicate varying levels of bioturbation. Assuming surface mounding in photographs of fine grained sediments indicates individual infaunal organisms (eg molluscs), up to hundreds of burrows per m^2 are evidenced in Photo P& (Plate E10). Infaunal activity in coarse sediments such as those in Photos 8SCI 4W&X (Plate E7) is difficult to estimate but burrow density may be in the thousands per m^2 .

Iceberg Turbation

Iceberg turbation is the process whereby icebergs plough large scours in the bottom sediments. Iceberg scouring intensity varies inversely with water depth, with maximum scouring occurring between 300 - 400 m (Dowdeswell *et al.* 1993). O'Brien (1992) interprets the jagged, small scale, randomly oriented

features seen in his provinces 1, 3, 4, & 7 as iceberg gouges and notes that shallow parts of the continental shelf are not cut by these features. O'Brien (1992) notes that these shallow areas are surrounded by "knick point and shoal" features (Figure 6.21) originally recognised by Barnes *et al.* (1987).

Iceberg scouring is reported to depths to 500 m on the continental shelf of in the Weddell Sea (Barnes and Lein 1988), to 550 m in the Scoresby Sund (Dowdeswell *et al.* 1993, 1994), and 720 m in Prydz Bay (O'Brien 1992). These extreme depths are explained by the rolling of tabular icebergs which may increase the keel depth of a berg by 50% (Lewis and Bennet 1984). Thus the 400 m (Budd *et al.* 1982) draft of the Amery Ice Shelf might produce, after rolling, iceberg drafts of up to 600 m. O'Brien (1992) explains the extra 120 m depth of gouging as occurring during lower sea levels.

Iceberg scours identified from side-scan sonar are continuous over thousands of metres (Dowdeswell *et al.* 1993). Scours on the east Greenland continental shelf are up to 10 m from crest to trough and with widths up to 20 m or more (Dowdeswell *et al.* 1993). Scours in Prydz Bay are of roughly the same dimensions and width according to the echosounder traces previously discussed.

Iceberg turbation can be identified by the resultant formation of low angle faults extending up to 6 m below the scour incision surface, large displacements, sub-horizontal thrust faults, high angle faults and large berms parallel to the scour. In addition, disharmonic, disarticulated folds with a well developed fracture cleavage, slickensided surfaces may be in evidence (Woodworth-Lynas and Guigné 1990).

Figure 6.4 contains a feature which may be an iceberg scour. The sediment surface which is flat lying in the top of the photo, slopes very steeply downwards at the bottom of the photograph. In cores, traces of the structural features mentioned above may provide evidence of iceberg turbation. Iceberg turbation shows up on echosounder and seismic traces (O'Brien 1992, fig. 7).

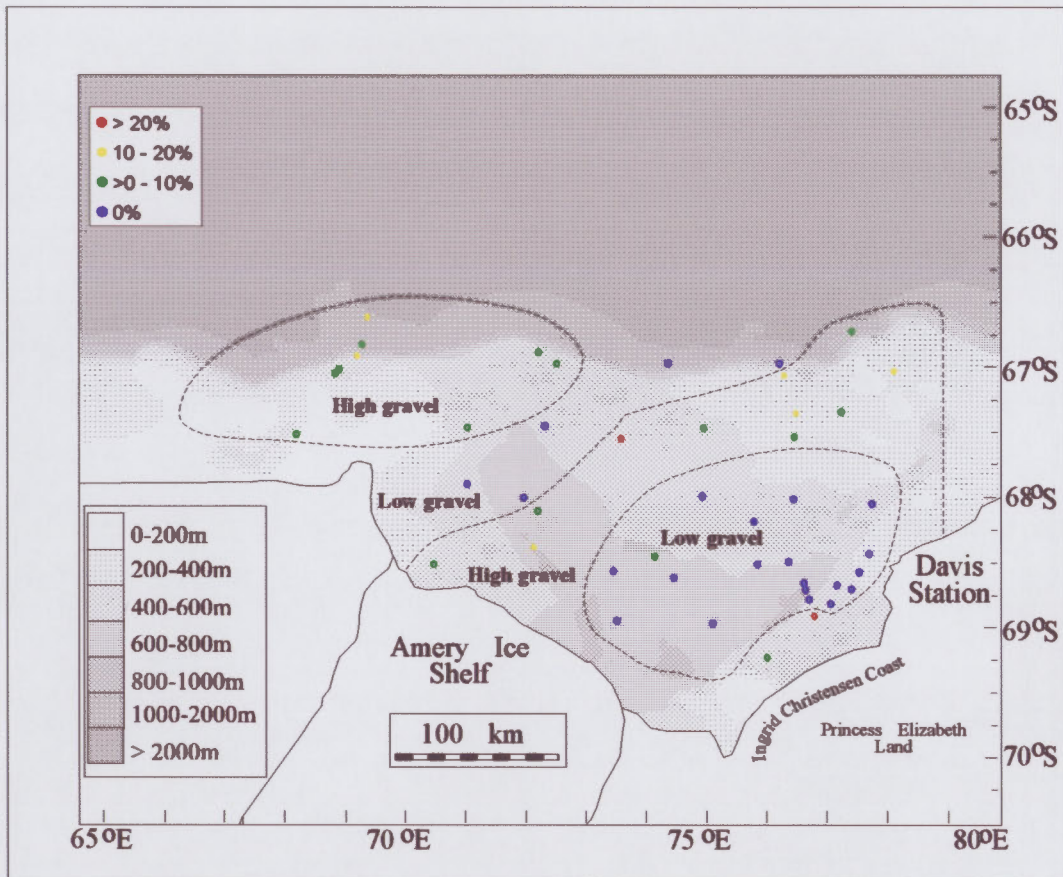


Figure 6.20 - Gravel in Prydz Bay sediments

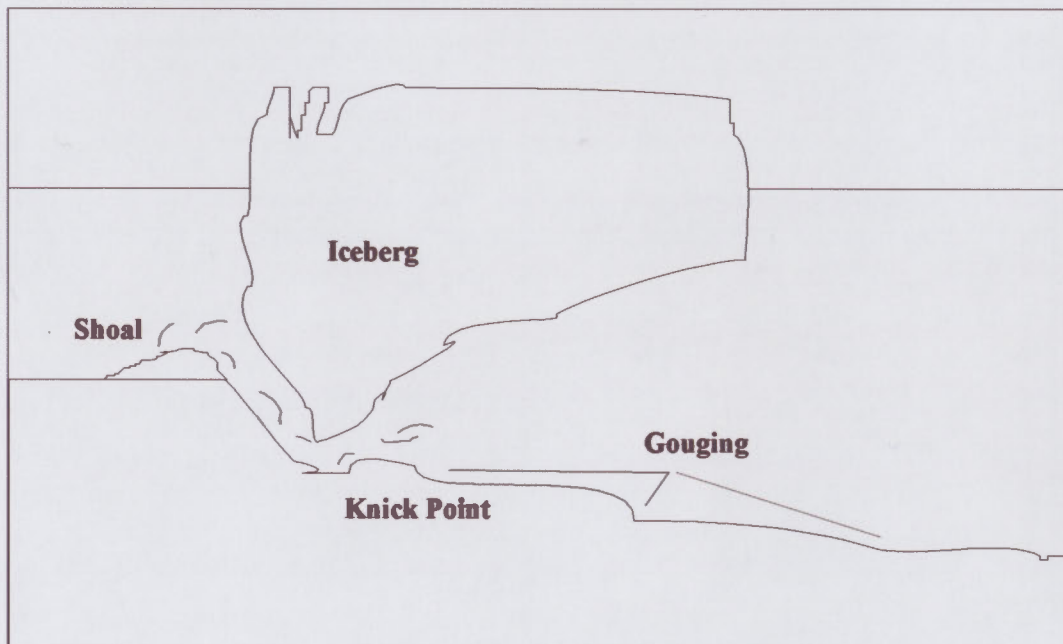


Figure 6.21 - Knick point and shoal formation by grounding icebergs.

Current reworking of sediments

The prevailing oceanographic regime in an area has an impact on the sediments which are deposited there, either by preventing certain size fractions from being deposited or by changing the sediments (grain size, diagenesis, dissolution, and mineralogy) after they have been deposited by other means. Current velocity data in Prydz Bay are restricted to those reported by Hodgkinson *et al.* (1988) but some research has been conducted into the oceanographic regime in Prydz Bay (Smith *et al.* 1984). The major feature of southern ocean circulation is the Antarctic Circumpolar Current (ACC) or "West Wind Drift" which is driven by the prevailing westerly winds (Pickard and Emery 1990) and which flows clockwise round the continent (Foldvik and Gammelsrød 1988). In the vicinity of Prydz Bay this current is located to seaward of the 2000 m isobath (Gordon and Molinelli 1982, Smith *et al.* 1984). Even though this current does not directly impinge upon the study area, it affects the currents within Prydz Bay.

Currents within Prydz Bay are dominated by a large cyclonic gyre which appears to be fed by a south trending current in the east of Prydz Bay which probably breaks off the ACC and flows into Prydz Bay (Smith *et al.* 1984, Franklin 1991). This water joins the east wind drift and flows westwards past the front of the Amery Ice Shelf where it mixes with meltwater from the iceshelf and then flows north along the coast of Mac. Robertson Land, rounding Cape Darnley and then westwards along the Mac Robertson Shelf into the vicinity of Mawson Station where it flows north and rejoins the ACC (Figure 6.22).

Bottom currents in Prydz Bay are mainly driven by three processes. Firstly, katabatic winds blowing off the continent drive water offshore and cause upwelling along the coast. Coriolis forces cause the waters to flow from east to west along the coast affecting the shallow coastal waters. Secondly, tidal forces cause ebb and flow currents which affect shallow waters. Finally, geostrophic currents operate at all levels of the water column. These currents are driven

by the differential distribution of mass in the ocean “which can be thought of as pressure gradients” (Pickard and Emery 1990). The distribution of mass in the oceans is related to the density of the water which varies vertically and horizontally. Density differences are caused by differences in the temperature and salinity of the water as well as the water depth. All of these factors result in horizontal pressure differences which will cause water flows from areas of high pressure to areas of low pressure. With the effect of Coriolis Force such flows can be inferred from the variation in temperature, depth, and salinity.

In the Weddell Sea, such currents have been reported as flowing parallel to bathymetric contours at speeds between 10-15 cm s⁻¹ (Pudsey 1992) where it produces a distinguishable sedimentary deposit described as Contourite facies (Grobe and Mackensen 1992) which consists of thin to medium beds containing horizontal laminae of mud and silty clay. These sediments represent the winnowed fines removed from sediments on the upper slope which, when reaching increased water depth are transported and deposited by weak bottom currents.

Fluctuations in current velocities are correlated with changes in the grain size distributions of the sediments such that in areas of high bottom current activity the particle size distribution will reflect the winnowing of the finer lighter fractions. Conversely, areas of low bottom current activity will not be so affected and may show grain size distributions relatively enriched in fines due to the transport of winnowed material and subsequent deposition in the low activity zone. Differences in statistical parameters of sediments can be used to discern this sort of current activity.

A useful indicator of the relative level of bottom current activity is the degree of skewness of the sediments (Figure 6.23), with coarse skewed sediments, those enriched in coarse material, representing high activity areas where bottom currents are winnowing fines out of bottom sediments, and fine skewed sediments representing low activity areas where those fines are settling out of suspension.

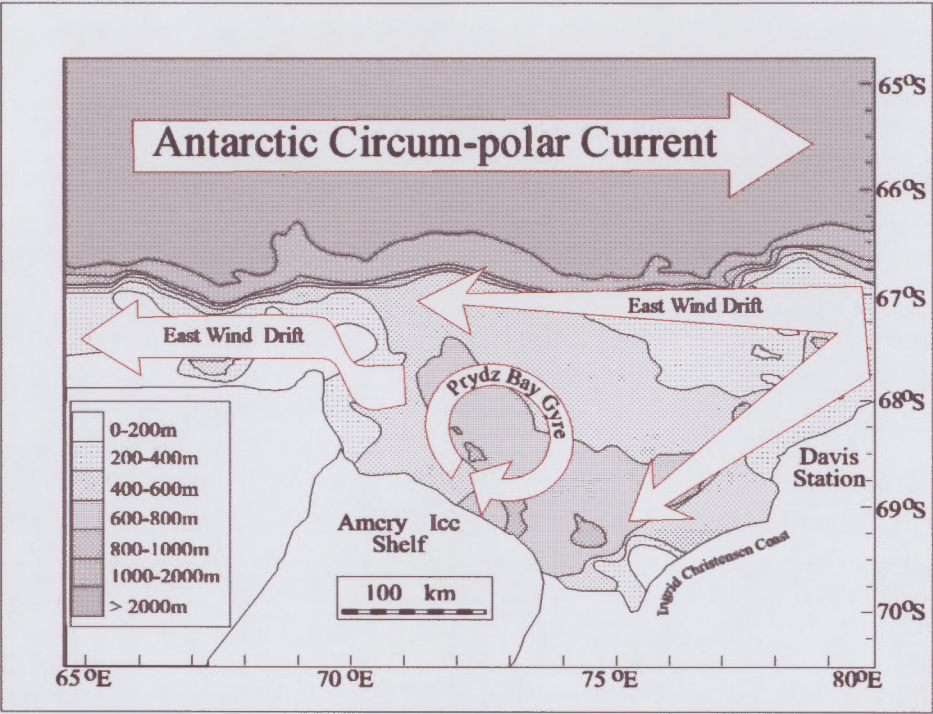


Figure 6.22 - Approximate position of major currents in Prydz Bay. [after Wong 1994]

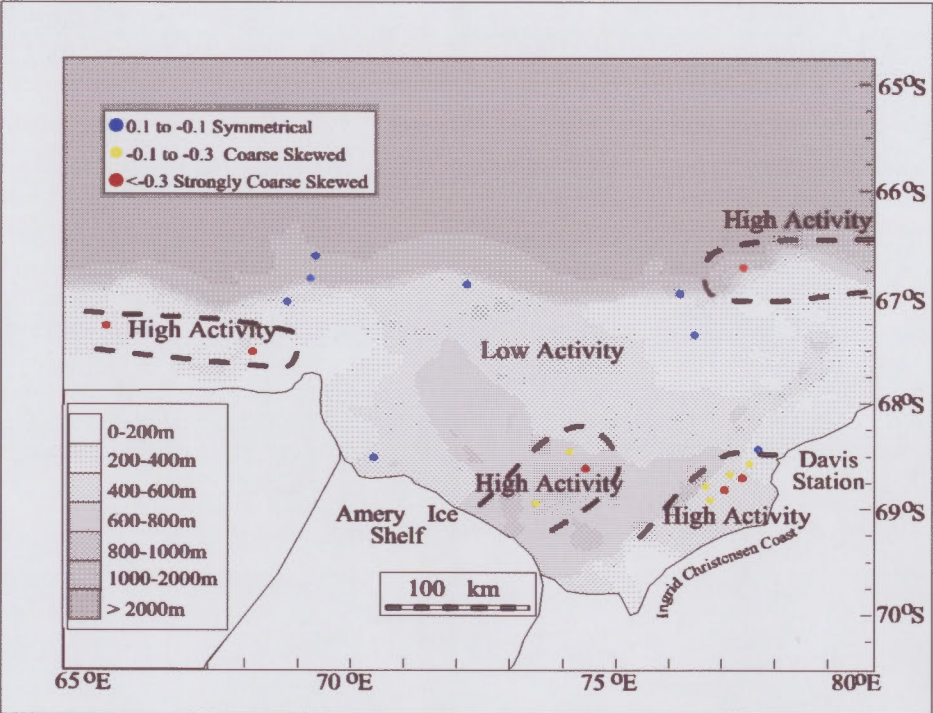


Figure 6.23 - Areas of current activity as indicated by skewness in surficial sediments.

The degree of skewness imparted on a sediment will also depend on the skewness of the sediments being supplied to the bay. In this instance the two samples collected which probably most closely resemble the terrigenous sediments being supplied by glacial processes are 93017 (Skewness -0.05 , or nearly symmetrical) from englacial debris in a grounded iceberg in the vicinity of the Vestfold Hills and 94004 (Skewness -0.16 , or coarse skewed) collected from a meltout deposit adjacent to an ice cliff in the eastern part of the Vestfold Hills. Statistical parameters from aeolian debris collected in the Vestfold Hills, the other probable source are in Table 6.2.

	Mean Grain Size (ϕ)	Standard Deviation (ϕ)	Skewness (ϕ)
Minimum value	0.68	0.61	-0.51
Average Value	1.25	0.87	0.08
Maximum Value	2.17	1.08	0.49

Table 6.2 – Ranges of values of grain size statistical parameters for aeolian material in the Vestfold Hills.

In Prydz Bay, the geographical variation in skewness values (Figure 6.23) is complex. Areas of high activity are found along the Ingrid Christensen Coast in the vicinity of Davis Station. These strongly coarse skewed sands may be the result of south west flowing currents being driven by offshore katabatic winds. Sea-ice rafting of sands from the Vestfold Hills may be important in this area and sea ice could be expected to travel 100's of kilometres before melting and releasing debris loads (Ian Allison – Personal Communication). In areas where aeolian debris is likely to be the main source of sediment, the sediments which have been subjected to winnowing will be those which display coarser skewness than 0.08ϕ .

In the samples collected close to Davis Station, the most proximal displays a nearly symmetrical skewness and probably indicates little or no current effect

on the debris supplied. Samples to the south west are coarse to strongly coarse skewed. Such a situation could be explained by either the preferential meltout of larger clasts in the early part of the sea-ice's travel to the south west or to the differential transport of fines by currents acting on debris settling through the water column. It is likely that both of these processes occur, and more sampling should be able to clarify these processes.

Other areas of high activity are represented by coarse skewed sediments off the front of the Amery ice Shelf. Water mass distributions and interactions are complex in this area and conductivity, temperature and depth data suggest that a number of currents meet and water masses mix at about 250 m to 500 m water depth (Franklin 1991). While the strength of these currents is unknown, the presence of the Prydz Bay Cyclonic Gyre is reasonably well established (Smith *et al.* 1984, Woehler and Williams 1987, Woehler *et al.* 1987) and, while it is difficult to quantify what effect they may have on sediments settling through the water column, there will be some effect.

This gyre could be expected to concentrate icebergs in its centre, quiet water zone. If so they would remain there and deposit coarse grained sediments below as the entrained material melts out. As all of the sediments below the centre of the gyre are relatively fine grained (M, sM, mS, cpmD, and cpiD) the bergs either do not concentrate or debris does not melt out. Both of these explanations are probably valid, with meltout already established as less important than other processes. Whether bergs concentrate in the centre of the gyre is not known but offshore winds would probably push bergs that were not entrained in ocean currents.

The final area of intense current activity as indicated by the skewness of the sediments is along the shelf break along the Mac. Robertson Shelf and in the north east of Prydz Bay. Currents in these areas have been measured by Hodgkinson *et al.* (1988, 1991 a & b) (between 10 and 25 cm s⁻¹) and these data are supported by iceberg drift tracks further to the east between 90 and 120°E (Hamley and Budd 1986) which show icebergs drifting westwards along

the continental shelf, driven by the east wind drift and then, further off the coast, eastwards in the ACC (west wind drift).

Another indicator of current activity is mean grain size (Figure 6.24). Geographic variability in this parameter supports the skewness data. While sediments tend to fine with increasing distance from the debris source, the variation in Prydz Bay is complex. Observed differences in mean grain size must again be considered separately with respect to the probable source material. Near Davis Station, where the source is probably sea-ice rafted aeolian sands, the mean grain size of the sand fraction fines with increasing distance from the Vestfold Hills. This is due to the preferential deposition of coarser grains early in the transport history.

The same is probably true for sands off the front of the Amery Ice Shelf and the Mac. Robertson Shelf where the transport mechanism is probably not sea-ice but iceberg rafting from icebergs in the Prydz Bay Cyclonic Gyre. The coarse sands on the shelf edge and shelf slope probably represent a lag deposit in which winnowing has been sufficient to overprint the depositional signal. In areas where iceberg and, to a lesser degree, biological turbation are important, mean grain size may also be affected by the mixing of very poorly sorted and probably coarser underlying subglacial tills, although this cannot be determined with this data set.

Distinguishing between these two patterns, that is the variation of mean grain size with distance from the source (a depositional signal) and the variation of parameters by current action (a reworking signal) can be achieved to some degree, by the degree of sorting of sediments. In general, the degree of sorting (standard deviation) of the grain size distributions is more strongly affected by current action; by the size sorting of sediments during bedload transport and the sorting of sediments during settling of debris through the water column. In the later instance, lateral movement of debris during settling will separate different fractions on the sea floor and result in higher degrees of sorting.

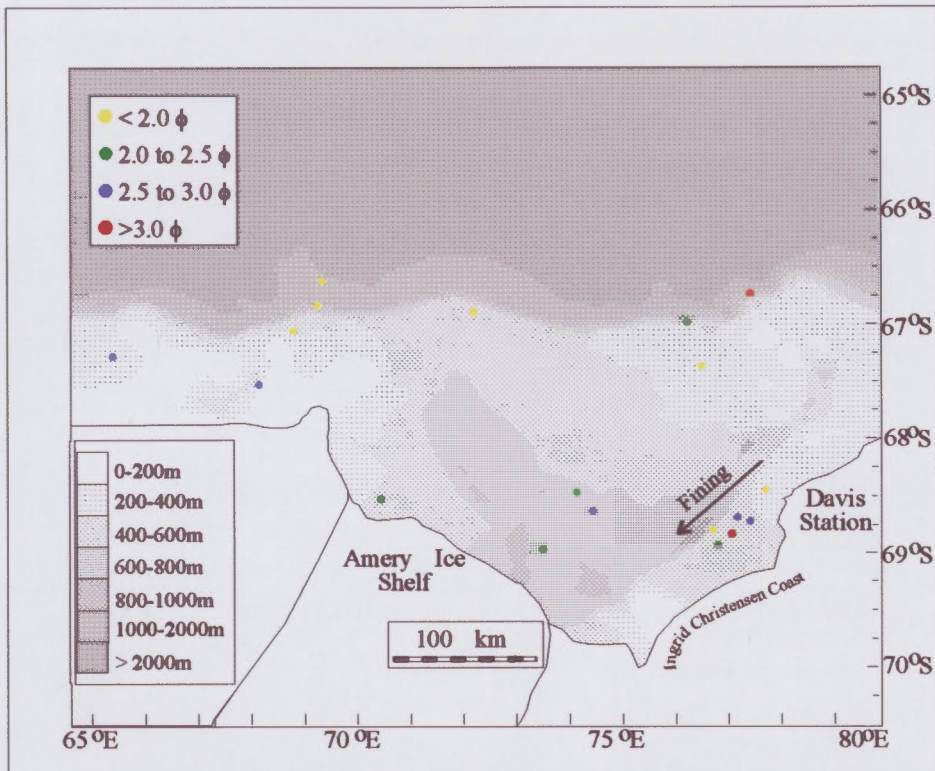


Figure 6.24 - Geographic variation in the mean grain size of the sand fraction of Prydz Bay surficial sediments. Value for englacial sample, 93017 is 1.58 ϕ , for meltout sample 94004 is 1.98 ϕ

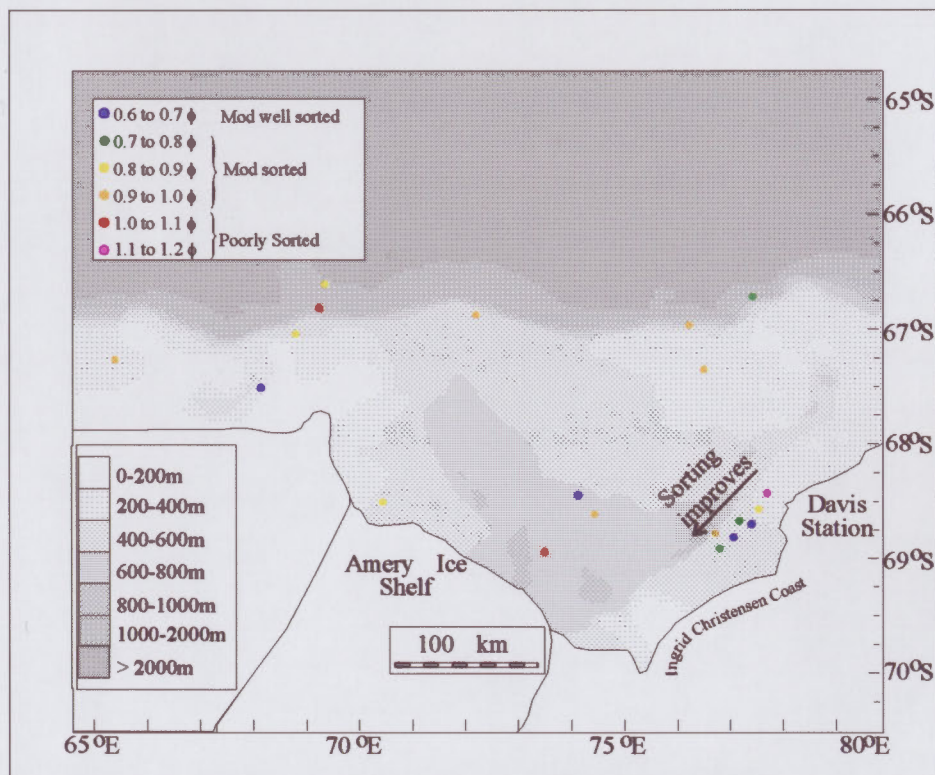


Figure 6.25 - Geographic variation in sorting of sand fraction of sediments from Prydz Bay. Englacial sample (93017) - 1.42 ϕ , meltout sample (94004) - 1.25 ϕ

In the vicinity of Davis Station (Figure 6.25), sediments are generally better sorted than those from other areas and sorting improves in the “downstream” samples suggesting that there is some current action which affects the moderately sorted source material. The high overall sorting of these sands supports an aeolian source. Conversely, sands offshore from the Amery Ice Shelf are disparate with standard deviations from poorly sorted to moderately well sorted, which may indicate that current action is not a dominant process here. Shelf edge and slope sites are problematic, with poorly sorted sediments displaying near symmetrical skewness and generally coarse mean grain size.

In summary, oceanographic influences in Prydz Bay seem to constitute three regimes. Near to Davis Station, the sediments, which are probably sourced from aeolian debris deposited in and on sea-ice, generally display better sorting, finer mean grain size, and coarser skewness. This is consistent with gradual meltout of debris laden sea-ice as that ice is transported to the south west by the east wind drift. Shelf edge and slope sediments are generally coarse grained, coarse skewed to symmetrical, and poorly sorted.

This pattern may indicate east wind drift currents winnowing fines and leaving lag deposits of coarse sands which are strongly coarse skewed. Those with symmetrical distributions are found in deeper waters > 800 m where currents may not be as strong. The samples off the Amery Ice Shelf consist of medium grained, poorly to moderately well sorted, and coarse skewed sands which are probably iceberg rafted sediments, with the coarse skewness due to separation of sediment fractions during the settling of debris through the water column.

Summary of Prydz Bay sedimentary processes.

Prydz Bay sedimentary processes (Figure 6.26) include the glacial entrainment and transport of debris by icebergs. Debris melts out of the glacial ice and settles through the water column to be deposited in the surficial sediments. In some places where there are significant ocean currents, the settling debris is

transported laterally, causing the fine material to be separated from the coarse material. In the vicinity of Davis Station and the Ingrid Christensen Coast debris from the Vestfold Hills is entrained by the winds and deposited on the top of sea-ice. When the sea-ice breaks out it is transported to the south west by the Prydz Bay gyre for up to 100's of kilometres and, as the sea-ice melts the sediment is introduced into the water column. Where bottom currents are active, fine material is winnowed from surficial sediments and transported until the current strength drops sufficiently for the fines to settle out of the water. In areas between ~ 200 and 720 m water depth, iceberg turbation mixes surficial sediments with underlying pre-Holocene glacial and/or glaciomarine sediments and bioturbation is probably ubiquitous. Finally isostatic uplift of ice free areas makes marine sediments available for erosion, modification, and reworking.

FACIES OF PRYDZ BAY

Pre Holocene Geology

The processes discussed are producing two major facies today (Figure 6.27); Siliceous Mud and Ooze (Facies SMO) and Iceberg turbate facies (Facies I), Rocks and sediments that were emplaced before the Holocene are the Relict Till (Facies T), and the Crystalline Basement. In some areas, calcareous sediments are also being deposited

Crystalline basement.

“Basement” underlies all other facies throughout the study area. It outcrops onshore in the Antarctic oases such as the Vestfold Hills, the Rauer Islands, the Larsemann Hills along the coastline, and in the various massifs and mountain ranges like the Prince Charles Mountains inland. There are areas where basement outcrops on the seafloor but more importantly, it can be seen to underlie surface sediments in areas of Prydz Bay, particularly adjacent to the Ingrid Christensen Coast.

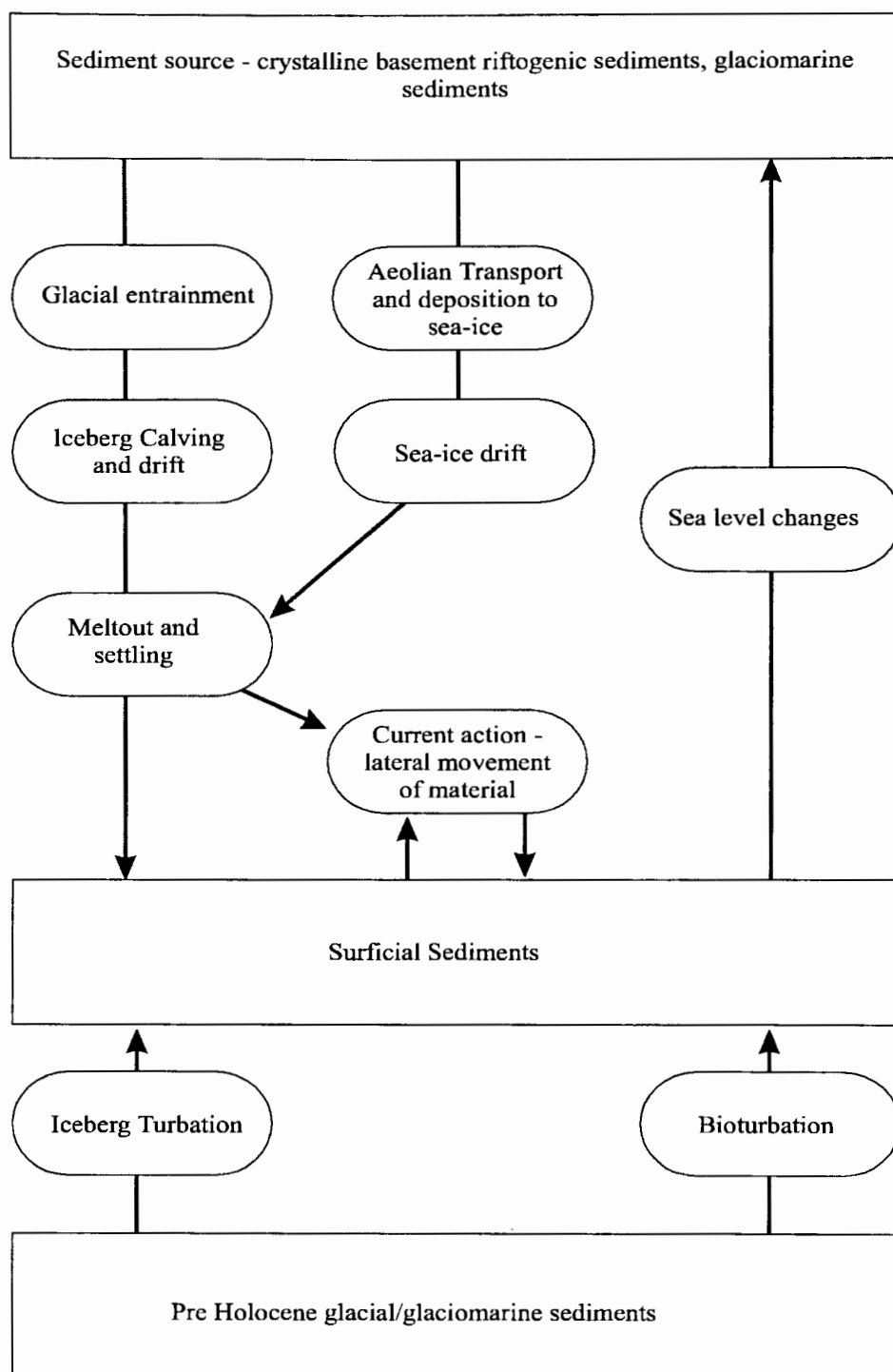


Figure 6.26 - Summary of processes affecting sedimentation in Prydz Bay.

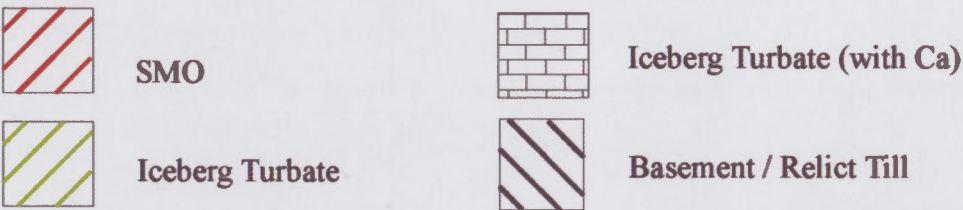
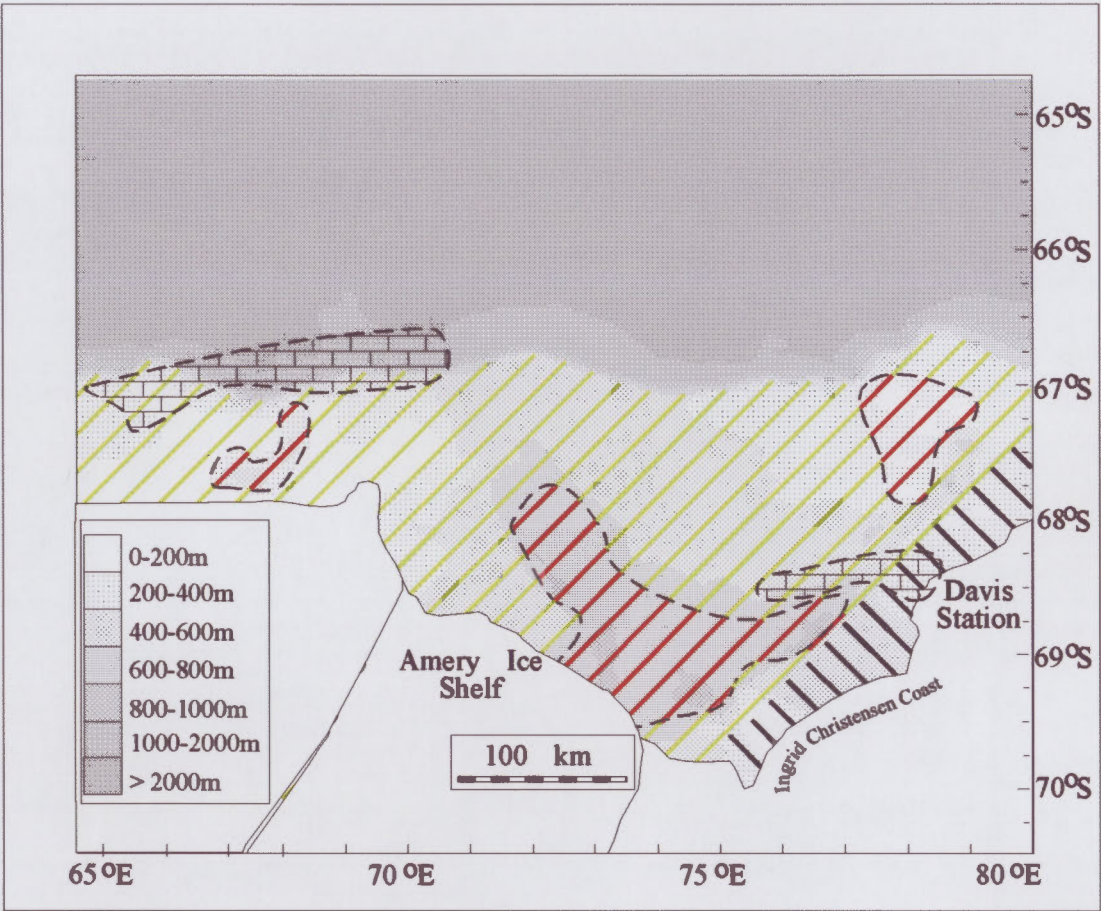


Figure 6.27 - Facies distribution in Prydz Bay.

“Basement” may also include younger sediments near Beaver Lake in the Prince Charles Mountains which are part of the Permian and Triassic Amery Group. These sediments are composed of conglomerate, gritstone, siltstone and shale; rhythmically bedded sandstone, siltstone, claystone, and coal; and arkose, coarse to fine sandstone, and claystone. Diatom ooze and diatomaceous mud with a minor ice rafted component” overlie Miocene to Quaternary massive to stratified diamictite. None of these sediments, however, are encountered in the process of this study.

Basement is identifiable at all scales by the lithology: highly deformed crystalline metasediments and younger riftogenic sediments and volcanics.

Facies T

Relict Till (Facies T) is composed of Miocene to Quaternary glacial and glaciomarine sediments and overlies crystalline basement through much of the study area, forming the top few hundred metres of the sedimentary pile in the Lambert Graben. It is composed of a mixture of all grain sizes from clay to boulders, and displays the very poor to poor sorting typical of glacial sediments. Clast lithology is variable but mainly represents the range of lithologies found in the basement rocks. It is generally moderately to highly indurated.

Facies T is overlain by Facies SMO and Facies I. It outcrops in areas adjacent to the Ingrid Christensen Coastline.

Holocene facies

Facies SMO

Siliceous mud and ooze (Facies SMO) is composed of sediments ranging from muds to clast poor muddy diamicts with a high to very high component of biogenic silica. It contains dropstones up to boulder size and lenses of coarse sediments. Dropstones are ice rafted glacially eroded continental rocks of

basement and Facies T. This facies forms massive, thick beds or elongate lenses and fill in structural depressions.

On a macro scale, Facies SMO displays a smooth surface in echosounder profiles and forms horizontal to sub-horizontal infills in the underlying topography. On the bed scale, this translates to sheets (inferred from the mode of deposition - ice rafting), stringers (seen as lenticular infills in the Svenner Channel in cross-section which can then be inferred to join up with similar ones in adjacent transects), and low angle drapes (seen in echosounder profiles). Bioturbation is ubiquitous and on a grain scale, the sediments are fine grained, the sand fraction poorly sorted, and grains angular to sub-rounded. The diagnostic features of this facies are the low gravel content and the high biogenic silica content.

Facies SMO is bounded below by the indurated glacial sediments of Facies T or basement. It overlies or underlies, or grades laterally into, sediments of Facies I. It is interpreted as being medial to distal glaciomarine sediments deposited by the concurrent deposition of vertically accumulated biogenic silica and ice rafted englacial sediments.

Facies I

Iceberg turbate (Facies I) contains all grain sizes from boulders to fine clays, terrigenous and biogenic material (carbonate and silica). The matrix is fine grained silts and clays and biogenic silica. Clastic material is dominated by breakdown products of Proterozoic metasediments and some younger riftogenic sediments and mafic volcanics. Facies I forms massive beds up to 10 m thick with no signs of meso-scale structure but large scale structures such as small (10 m deep, 40 m wide) troughs can be seen in echosounder profiles and these features can also be seen in side scan images as long, randomly oriented gouges in the surface sediments.

Facies I overlies Facies T and basement and grades laterally into Facies SMO. In places it may be overlain by drapes of facies SMO. Facies I can be identified

at the macro scale by the irregular, angular appearance of its surface in echosounder profiles, with apparent slumping and soft sediment deformation (GC30, Figure F.1). Inferred within-bed structures include evidence of soft sediment deformation such as slump surfaces, slickensides, normal faulting etc. Grain scale details include the presence of clasts of indurated sediments from Facies I. The diagnostic feature of Facies I is the surface appearance in echosounder profile, on the meso-scale and the presence of clasts of indurated sediments at smaller scales.

Facies I is interpreted as a reworked admixture of facies SMO material and the underlying Facies T. As this mixing is achieved by the deep turbation of sediments by iceberg scouring, this facies is restricted to depths of between 200 and 720 m.

Calcareous Sediments

While not considered a separate facies, calcareous sediments are being deposited in shallow areas of Prydz Bay, mainly adjacent to the Ingrid Christensen Coast and on the shelf edge and shelf break of the eastern part of the Mac.Robertson Shelf.

PRYDZ BAY SEDIMENT BUDGET

Sedimentary Budget.

The sedimentary budget for Prydz Bay will be made by calculating the mass of sediment (M in Mt a^{-1}) deposited, based on the following equation,

$$M = \frac{r \cdot a \cdot \rho \cdot (1 - p)}{10^3} \quad \text{Equation 6.3}$$

Where r is the rate of accumulation in cm a^{-1} , a is the area over which accumulation is occurring, ρ is the density of the sediment, and p is the porosity of the sediment as a fraction of 1.

Sediment budget caveats

The rate of sedimentation is based on an average rate from the three cores described.

The ^{14}C ages from the cores used to determine the sedimentation rate are uncorrected and the reservoir effect is assumed to be constant throughout. The sedimentation rate is based on only these cores and their accuracy is probably compromised by this.

The proportion of biogenic carbonate in Prydz Bay sediments is difficult to quantify as there were insufficient determinations of sediments of all types to provide a reliable result.

Some areas of Prydz Bay may be erosional, and while there is evidence of winnowing this study assumes that all areas of Prydz Bay are depositional and the sampling program is assumed to represent all sedimentary environments.

Sediment porosity is assumed to be 0.4 (Dr. Peter Harris, personal communication).

Rates of accumulation of sediment

^{14}C dating on three cores GC03, GC29, and GC30 (Figure F.1), provided sedimentation rates of 0.11 mm a^{-1} , 0.16 mm a^{-1} , 0.08 mm a^{-1} respectively. GC 03 was a mainly sandy core collected to the north of the study area and may not be representative of the processes occurring within Prydz Bay. GC 29 is a 352 cm core of Facies SMO that appears to indicate a record of sedimentation without erosional periods for the last 24000 a. The surface sediments here were dated at $\sim 2500\text{a}$. Core GC 30 is a 150 cm core of structureless to whispily laminated diamict. The surface sediments are dated at $\sim 6500 \text{ a}$ and it is likely that this surface has been eroded or strongly turbated. In deciding which value to use as the sedimentation rate, the following considerations are made. The lens of SMO (GC 29) seems to represent a hydrological sediment trap so the value of 0.16 mm a^{-1} is considered a maximum possible rate. The

value of 0.08 mm a^{-1} for GC 30 is probably closer to the true rate of sedimentation but, because of the lack of significant biogenic silica and the possibility of erosion indicated by the very old age of the surface sediments, probably underestimates the sedimentation rate. The value for GC 03, although representing the shelf sedimentation rate is between these two values and is probably closer to the true rate and is therefore assumed to apply throughout the bay.

Area over which the accumulation is taking place

Prydz Bay is roughly triangular with dimensions of 400 km by 300 km. This gives an approximate area of $60\,000 \text{ km}^2$.

Rate of Sediment Accumulation.

Substituting values into Equation 6.1 (Table 6.3) the total amount of sediment being deposited in Prydz Bay is 10.33 Mt a^{-1} , a very low value supporting other authors' (Quilty 1985, Barron *et al.* 1991) assertions that the sediment supply to Prydz Bay is impoverished.

Fraction	Area (km^2)	% of whole	Sedimentary Rate (mm a^{-1})	Density (g cm^{-3})	Porosity	Mass (Mt a^{-1})
Gravel	60 000	6.20	0.11	2.65^{\dagger}	0.4	0.65
Sand	60 000	23.60	0.11	2.26^{\dagger}	0.4	2.48
Mud	60 000	56.77	0.11	2.26^{\dagger}	0.4	5.96
Total Terrigenous						9.08
Silica	60 000	8.61	0.11	2.15^{\ddagger}	0.4	0.73
Calcite	60 000	4.82	0.11	2.70^{\S}	0.4	0.52
Total Biogenic						1.25
Total						10.33

[†] Density of Quartz
[‡] Density of Opal
[§] Density of Calcite

Table 6.3 – Mass of sediments accumulating in Prydz Bay

With such an apparently small amount of sediment being deposited in Prydz Bay, it may be possible to identify the major sources of the terrigenous component. The possible sources are the Lambert Glacier / Amery ice Shelf, the glaciers draining the Ingrid Christensen Coast (Sørsdal Glacier, Ranvik Glacier, Polar Record Glacier, etc), icebergs from the West Ice Shelf brought into the bay in the east wind drift, and finally, the sea-ice rafted, aeolian deposited material from offshore the Vestfold Hills and other continental oases.

Sediment is incorporated into glacial ice in the Lambert Glacier and the glaciers fringing the Ingrid Christensen Coastline. Most of this sediment is incorporated in the basal ice of these glaciers and transported to the sea (Hambrey 1991). The amount of sediment typically found in glacial ice is dependent on, *inter alia*, thermal regime (dry or wet based), thickness of the englacial debris zone, sediment concentration in the englacial debris zone, and the prevalence of exposed areas of rock around the margins of the glacial catchment. Dowdeswell and Murray (1990) summarised the variability in these parameters in respect to four glacial types indicating that dry-based glaciers have debris rich basal layers up to 10 m thick and containing up to 50% (by volume) entrained debris.

Values for the englacial debris content of glacial ice of 0.001% (Dowdeswell 1986) were used to model the sedimentation rates from icebergs, and included an assumption that the basal ice thickness for high polar glaciers is 2 m. Actual values of englacial sediment in Antarctic glacial ice were determined by Anderson *et al.* (1980) in observations of a number of icebergs suspected of originating in the Mertz and Ninnis Glaciers. A sample from a debris rich layer of 8 m thickness yielded 0.2 g of sediment from a 5 cm³ sample of ice (4% by mass) and another yielded 5% by mass debris from a debris rich layer. Other samples yielded similar amounts but the results represent amounts in dirty bands and not amounts with respect to the entire thickness of glacial ice.

During the collection phase of this study, one sample, 93017, of glacial ice from a debris rich layer was collected which contained 32.17 g of sediment in

3120 cm³ (0.01 g ml⁻¹) of glacial ice. The debris layer was ~ 2 m in thickness and the sediment load was therefore about 1% of the ice mass. This figure is within the bounds of the variability discussed by Dowdeswell and Murray (1990), although low in the range, and is probably lower than the actual figure because, for instance, there were several large clasts visible in the iceberg which were not sampled (Figure 4.1).

The rate at which ice is being discharged into Prydz Bay has been estimated at 11 Gt a⁻¹ at the grounding line of the Lambert Glacier (Hambrey 1991). There is considerable basal melting near the grounding line such that some 40% of the ice mass is lost. One third of this volume is replaced by basal freeze-on of saline ice. This process occurs very close to the grounding line and probably "locks" in any englacial sediment that has not already melted out so that it is transported well away from the ice shelf. Hambrey (1991) suggested, on the basis of known drift tracks, that these icebergs leave the Prydz Bay area before contributing any significant amount of sediment to the basin. The Lambert Glacier is more than 600 m thick in the vicinity of the grounding line but thins to around 270 m thick at the calving line.

The amount of sediment being supplied to Prydz Bay via glacial erosion and ice rafting is in the order of 9 Mt a⁻¹ (Table 6.3). As many workers have conducted investigations into the amount of sediment carried in glacial ice, it is now possible to determine the amount of sediment that could be expected to be supplied to Prydz Bay via the Amery Ice Shelf using the following equation.

$$m_s = \frac{w_{bi} \cdot \%sed_{bi} \cdot v_i \cdot \rho}{100 \cdot w_{gi}} \quad \text{Equation 6.4}$$

Where m_s is the mass of sediment in a volume of glacial ice measured in Mt a⁻¹, w_{bi} is the width of the basal ice layer in m, $\%sed_{bi}$ is the percentage of sediment in basal ice by volume, v_i is the volume of ice, ρ is the density of the sediment in g (cm³)⁻¹, and w_{gi} is the width of the glacial ice in metres. Using this equation

and some typical figures for each of the variables the following theoretical sediment masses are determined (Table 6.4).

Variable	Low value	Median Value	High Value
v_i	11 Gt a ⁻¹	30 Gt a ⁻¹	60 Gt a ⁻¹
ρ	2.65 g cm ⁻³	2.65 g cm ⁻³	2.65 g cm ⁻³
%sed _{bi}	10	30	50
w_{bi}	10 m	10 m	10 m
w_{gi}	600 m	600 m	600 m
m_s	41.43 Mt a ⁻¹	1130 Mt a ⁻¹	339 Mt a ⁻¹

Table 6.4 – High, median, and low values for the calculation of sediment mass (Mt a⁻¹) expected to be transported via the Amery Ice Shelf

Comparing the lowest expected sediment transport values in the Lambert Glacier with the (probably) high values of sediment mass accumulation in Prydz Bay the disparity between the two figures is about a factor of about 5. This is probably sufficient to conclude that most Prydz Bay sediments are not supplied via the Lambert Glacier but from a smaller glacial source.

There are a number of hypotheses which can explain the disparity between the amount of sediment being supplied to Prydz Bay and the amount that would be expected were the main source the Lambert Glacier. Firstly, the sediment may be being supplied by another glacial source, one approximately $\frac{1}{5}$ th of the size of the Lambert Glacier. The prime candidates for this are the glaciers draining the Ingrid Christensen Coast (the Sørsdal Glacier, Polar Record Glacier, and the Ranvik Glacier) and areas of coastal ice cliffs. The implication that would have for palaeoclimate reconstruction is that changes in sedimentary pattern or regime within the Holocene reflect changes to the coastal fringing glaciers and not changes to the Lambert Glacier. Because the Lambert Glacier catchment is much larger than the coastal catchment, then any glacial fluctuations in the Holocene inferred from these sedimentary changes probably represent small

scale climatic change as opposed to the large scale changes that would be required to cause changes in the Lambert Glacier system.

A second hypothesis is that the material entrained in Lambert Glacier ice is being released at a very low rate. This is supported by the argument already discussed, that melt rates of icebergs in Prydz Bay are probably very low due to the low ($< 0^{\circ}\text{C}$) water temperatures in Prydz Bay for most of the year. As previously discussed, the englacial material in Lambert Glacial ice that does not melt out close to the grounding line is "locked in" by basal freeze on of saline ice which occurs close to the grounding line of the Amery Ice Shelf and may result in saline ice of up to 200 m thickness on the bottom of the ice shelf. Whether this has any great affect on the prevention of the melt out of debris is unclear, however. Saline ice will melt at a much lower temperature than the freshwater ice and it may melt away very quickly in the waters of Prydz Bay, exposing the debris rich glacial ice.

A third hypothesis is that the debris is liberated from icebergs which originate outside of Prydz Bay, for instance, the West Ice Shelf, and are carried into the bay in the east wind drift, but no studies have tracked such icebergs.

Finally, the sediments may originate in onshore oases and be transported into Prydz Bay via aeolian deposition to sea-ice and subsequent rafting of that sea-ice. In Chapter 4, offshore transport of up to 0.6 Mt a^{-1} of debris was described occurring from the Vestfold Hills. If it is assumed that the rate of such transport represents aeolian transport from all continental oases then oases along the Ingrid Christensen Coastline (Vestfold Hills, Rauer Group, Larsemann Hills) probably supply approximately 1 Mt a^{-1} of debris or about 11% of the sediment accumulating in Prydz Bay. This input is probably most important proximal to the Ingrid Christensen Coastline.

The supply of material to Prydz Bay is probably attributable to all of these sources, with sea ice rafted aeolian debris important proximal to continental oases, and a mixture of all of the other sources throughout the rest of the bay.

The relative inputs from each source cannot, however, be discriminated with these data.

CHAPTER SUMMARY

The main two sediments which are currently being deposited in the offshore part of the study area, that is Prydz Bay proper and along the Ingrid Christensen Coastline in particular, are SMO and Iceberg Turbate. The distribution of these sediments is complex and dependent on a variety of processes.

SMO is a fine grained sediment with some larger particles and is deposited through the vertical accumulation of sediments melting out of icebergs and sea-ice, and the vertical accumulation of siliceous biogenic material composed mainly of diatom frustules, and the lateral accumulation of fines transported by bottom currents. Turbate is composed of clasts of all grain sizes from boulders to mud, is depauperate in biogenic silica when compared to SMO, and is formed by the turbation of SMO and underlying sub-glacial tills deposited during previous glacial maxima.

The processes which deposit the fine grained sediments found in SMO have little effect on the final distribution of SMO. Biogenic production of silica occurs throughout the bay. Diatom blooms occur along the Antarctic ice edge in the spring and follow the retreating ice edge during the spring and summer until it reaches the coast in mid summer. Over time biogenic silica is supplied to all areas of the bay in roughly equal amounts. As most of the diatom frustules are delivered to the sea-floor incorporated in the faecal pellets of zooplankton, there is probably relatively little lateral transport of this material by current action as the pellets behave hydro-dynamically as sand (Chris and Frakes 1971).

Iceberg Turbate is formed by the turbation of old sub-glacial tills and the sediment which is being deposited now. As such, turbate is found in all areas where icebergs are able to gouge the sea floor. Echosounder records show

turbation occurring in depths as great as 720 m but not all sediments shallower than 700 m are turbates. SMO is commonly found in shallower depths. In particular, many areas are protected by the submarine topography. Very large icebergs with very deep keels are grounded against submarine ridges and cannot pass until the keel depth is less than the depth of the topographic feature.